Frontispiece.

Photo by A. K. Lobeck.

Wave-cut bench and marine cliff bordering Desecheo Island, Porto Rico, recently elevated above sealevel.

SHORE PROCESSES

AND

SHORELINE DEVELOPMENT

BY

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PREFACE

THE present work was born of a need experienced by the author in connection with his shoreline studies. In the course of a critical examination of the arguments supposed by many to demonstrate a progressive subsidence of the Atlantic coast of North America within historic time, it developed that in respect to certain of these arguments agreement between students of the problem could not be reached because there was not sufficient agreement as to what features are normally characteristic of a stable coast, and what features are peculiar to coasts which are rising or subsiding. No work existed which combined with an extended analysis of the forces operating along the shore, a full and systematic discussion of the cycle of shoreline development and such further discussion of the modifying effects of changes of level as would enable one to differentiate stable, rising, and subsiding coasts.

It seemed necessary, therefore, to enquire somewhat fully into the fundamental principles of shore processes and shoreline development; for it would not be profitable to add to the already overburdened literature on changes of level another essay which should merely add quantitatively to the volume of evidence previously discussed by many earlier writers, and again assert as the correct interpretation of that evidence conclusions which some geologists and geographers accept and others reject. Profit could come from the study only in case the discussion of principles was such as to bring geologists and geographers into substantial agreement as to what shore features are, and what are not indicative of changes of level. Once this measure of agreement was reached, I could not doubt that a critical analysis of the arguments supposed to prove the progressive subsidence within historic time of the coast of southeastern Canada, the Atlantic coast of the United States, and certain other marginal areas of the continents, would demonstrate to the impartial and critical student the inadequacy of those arguments. I therefore set myself the task of bringing together the results of shoreline studies published in different languages, of analyzing and criticising the conclusions reached

where this might appear to be profitable, and of presenting a digest of those fundamental principles which should prove to be best established by the independent work of different students and best supported by my own field observations. At the same time I purposed to develop somewhat fully certain important aspects of the physiography of shorelines which have hitherto received little consideration.

The magnitude of the task proved to be greater than anticipated, partly because of the wide divergence of expert opinion regarding the manner in which shore processes operate, and partly because of the great volume and scattered distribution of the writings dealing with the subject. It was, indeed, the desire to relieve others who might have occasion to study shore processes and shoreline forms, of the burden of duplicating the work involved in my undertaking which first suggested to me the desirability of placing on record, in compact form for their use, the results of my enquiry, even where these results did not relate to the original problem of coastal subsidence. The present volume is the concrete product of this desire to render a service to my fellow students.

The engineer will find in the chapters on waves and currents a summary of the widely conflicting opinions and observations relating to those most puzzling forces with which he has to deal. In the later chapters he will also find, I hope, not a few discussions of shore forms and of the method of their development which will prove useful to him in his work on marine engineering structures. The dynamic geologist will find in the first chapters an extended account of two of the forces of nature with which he is much concerned, and in the remaining chapters abundant illustrations of the manner in which those forces operate near the margins of the lands. The geographer will be mainly concerned with the last seven chapters where the forms of the shoreline receive a systematic treatment which, if not adequate, is at least somewhat more detailed and complete than any hitherto attempted. Throughout the volume the reader will note that repeatedly conclusions reached and principles established are briefly applied to the problem of changes of level; and he will understand that this is the thread, appearing now and then, which is to connect parts of the present study with a later volume devoted exclusively to the much mooted question of coastal subsidence.

As a rule an *advance summary* precedes, and a brief *résumé* concludes the text of each chapter. This will enable engineer, geologist and geographer to determine in some measure the extent to which matters pertinent to their respective fields are discussed. A bibliography, arranged alphabetically according to authors and placed at the end of the volume, supplements the list of references given at the close of each chapter. Finally, an index of authors and an index of subjects are provided in the form which it is hoped will prove most serviceable to the reader.

In any attempt to give proper credit for the aid rendered by others during the preparation of this volume the writer is much embarrassed. The work of preparation has extended over several years, during which time a number of students, colleagues and friends have been most generous in rendering valuable assistance. It would be impossible to make specific acknowledgments to all of them, so great is the measure of my indebtedness. Special thanks are due to my cousin, Miss Laura Dale Johnson, for assuming the labor of reading the proofs and seeing the book through the press during my absence; to Miss Florrie Holzwasser of the Department of Geology of Barnard College, and others among my graduate students, for assistance in reviewing and abstracting the literature relating to the subject in hand; and to Dr. A. K. Lobeck for preparing the five block diagrams showing successive stages in the development of a shoreline of submergence. Acknowledgments should be made to "The Geographical Review," "Science," the "Bulletin of the Geological Society of America," and the "Journal of Geology" for the use of certain material originally published in their pages. Many of the observations recorded in this volume were made in the course of a special Shaler Memorial Investigation of the problem of coastal subsidence undertaken with the support of the Shaler Memorial Fund of Harvard University; and observations on the New Jersey coast were obtained in connection with a study in progress for the Geological Survey of New Jersey under the direction of Dr. H. B. Kümmel. It is a pleasure to express special obligations to Professor W. M. Davis for helpful criticism of the manuscript, and to acknowledge the debt which all physiographers owe to his studies of shoreline topography, which were the first to demonstrate the value of applying the idea of the cycle to the history of shore forms. To Professor Joseph Barrell, whose studies have to some extent paral-

leled certain of my own, I am indebted for many valuable suggestions and for his generous courtesy in giving to the manuscript a careful and critical reading, from the results of which I have greatly profited.

In conclusion it is but fair to acknowledge the author's keen appreciation of certain defects which the reader may discover in his perusal of the text. The volume goes to press under circumstances which absolutely prevent that careful attention to details which every work of this kind should receive. On entering the service of his country the writer was forced to choose between publishing his studies without the final supervision which he had hoped to give the proofs, and postponing publication indefinitely. In view of the uncertainties attending service in the zones of military operations, it has seemed wiser to allow the work to go to press, in the hope that the indulgent reader will not find the value of the volume materially affected by such errors in execution as the presence of the author alone could have prevented.

DOUGLAS WILSON JOHNSON.

On Board Troop Ship,
April 3, 1918.

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CHAPTER I

WATER WAVES

Advance Summary. — No adequate appreciation of the many problems presented by the shoreline can be gained until one is familiar with the work of waves and currents. The relative importance of these two forces in shaping the shore is a much disputed point; and the difficulties involved can best be set forth, and an attempt at their solution can best be made, if we review the essential characters of waves and currents with some care, and critically examine the manner in which each operates. We will first turn our attention to the phenomena of water waves of different types; then we will be in a position to discuss the work accomplished by such waves; after which currents and their work will be considered.

In this first chapter, after a note on the general scope of the present treatment of waves, there is presented to the reader a brief survey of the literature on the subject, which may be useful in showing the growth of our knowledge of waves since the time of Leonardo da Vinci. Attention is then directed to the two types of waves which are most effective in shore processes: the wave of oscillation and the wave of translation. In each case the origin and nature of the water movement are explained and the elements of wave form are described. The depths at which waves break on approaching a coast determine the position of certain shore forms, and therefore receive consideration. The factors affecting the height of waves are of vital interest to the engineer employed on harbor works or coast defenses, and to the student of shore forms produced under varied conditions of wave attack; hence these factors

are discussed with some fullness. A wave's capacity for destruction depends upon both its height and its length, and the velocities of certain waves vary with wave length according to definite laws. The lengths of waves and their velocities are therefore matters of importance to engineer and geologist, and the natural laws which govern them possess a fascinating interest for the laymen interested in one of the most impressive of Nature's destructive forces.

Earthquakes and explosion waves are comparatively rare phenomena, but their spectacular character, the popular interest which attaches to them, and the disasters for which they are responsible, entitle them to consideration in any treatise on waves. The great wave known as the "tide" is of importance to the student of shore processes only in connection with the currents which it produces, and is accordingly given scant space in Chapter I. A treatment of tidal currents will be found in Chapter III. Standing waves including the seiche, and the so-called "boundary waves" produced at the contact of liquids having different densities, are of theoretical rather than practical interest in the present connection, and are discussed very briefly.

Scope of Subject. — Water waves may be produced in a variety of ways. The bow of a vessel pushing through the water, or a strong wind blowing over the sea, or a rain-drop falling into it, will each produce waves; but in each case the waves are of essentially different character, and behave according to distinctly different laws. One of the waves generated by a submarine displacement of the earth's crust is similar to the wave pushed out from the bow of a moving ship, but is unlike those produced during a storm at sea. Waves which form when a fine wire is drawn through the water behave very differently from the ship's waves, but are like the ripples set in motion by a falling drop of water. The great wave known as the " tide " is a compound wave, combining some of the characteristics of the two groups of waves first mentioned. When wind-made waves break on a shallow shore they give rise to a new series of waves similar to those produced by the bow of a vessel. Such facts as these are sufficient to show that the subject of waves is an extremely complicated one. Opportunity for making direct observations of wave motion below the surface of water bodies in nature is very limited, while the

theoretical treatment of wave motion carries one into the realms of higher mathematics.

It is not within the scope of the present report to enter into a discussion of all the interesting series of water waves known to science. The beautiful and complicated wave pattern developed by a moving ship has little to do with the modelling of shore forms, and the reader who would follow that phase of the subject further is referred to Lord Kelvin's popular lecture "On Ship Waves"[1], the second chapter of J. A. Fleming's little volume on "Waves and Ripples"[2], and the twelfth chapter of Vaughan Cornish's book entitled "Waves of the Sea and other Water Waves"[3], in which latter place will be found exquisite photographic illustrations of ship waves. The phenomena of ripples are treated at some length by Fleming[4], and a more technical account is given by J. Scott Russell[5]. Other interesting forms of water waves are described at length by Russell[6] and Vaughan Cornish[7]. We must confine our discussion to those types of wave motion which have a significant effect upon the shore.

But even if we limit ourselves to a consideration of those waves of practical importance to the engineer and physiographer, our task is by no means an easy one. The literature of the subject is extensive and much of it highly technical in character. Different authorities employ different formulae in deriving some of the elements of wave motion, and the results they obtain agree neither with each other nor with the results obtained by experimentation. Airy commends the experimental work of J. Scott Russell as being the best ever done, but warns the reader against accepting that author's theoretical expressions, claiming that his own formulæ express the true relations and are verified by Russell's results[8]. Russell, in turn, demonstrates the inaccuracy of Airy's formulæ, and deplores the fact that the methods of investigation employed by that able authority should not have led him to better conclusions[9]. Hagen likewise opposes with some vigor certain of the suppositions made by Airy, while the experimental observations of Caligny and Russell disagree on important points. Krümmel has well expressed the present condition of the subject in the words: "In short analysis, observation and experiment are not yet in the desired agreement"[10]. Fortunately, a number of the disputed points are not of special importance to the student of shore forms, however much he

may be interested in the complex but beautiful laws which govern the motions of waves.

Literature.—Some of the principal sources of information upon which I have relied, and to which the student of waves is referred for elaborate discussions, may briefly be mentioned. Of historical interest are the work of Leonardo da Vinci, who in the latter part of the fifteenth century recognized many of the fundamental principles of wave motion, and advanced theories which are similar to those of modern investigators; and of Newton, who a century later gave us the first exact mathematical treatment of waves. Among more recent works the publications of Franz Gerstner, which appeared in the early years of the nineteenth century, are especially important. I have not seen the original papers of these authors, but their work is reviewed by the Weber brothers, Cialdi, Wheeler and others, in reports mentioned below.

In 1809 Bremontier's able essay entitled "Recherches sur le Mouvement des Ondes"[11] was published. This early report of experimental work on the laws of wave action and of observations on wave action in nature, contains the first effective demonstration of the power of waves to affect the bottom at considerable depths. The important volume of the two Weber brothers on "Wellenlehre auf Experimente gegrundet"[12], based on elaborate experimental studies and published in 1825, contains a review of practically everything written on waves from the time of Newton up to 1820, and adds much to the sum of previous knowledge on the subject. Six years later Emy's treatise "Du Mouvement des Ondes et des Travaux Hydrauliques Maritimes"[13] refuted Bremontier's conclusion that during wave movement the water particles rose and fell in vertical paths, substituted the more nearly correct opinion that the particles moved in vertical ellipses, and developed at great length the theory that a special type of "bottom wave" (flot de fond) was the principal cause of changes in the forms of the coast and of the destruction of maritime engineering structures. Emy does not appear to have been familiar with the work of the Weber brothers. J. Scott Russell's two reports on "Waves"[14], made to the British Association in 1837 and 1842–1843, present the results of admirable experimental work made under conditions more favorable than those attending the experiments of the Weber brothers, although

Russell directed his attention principally to the waves of translation. In reading Russell's reports the student must guard against misapprehension arising from the fact that the text references to plate numbers and to the lettering of illustrations are full of errors. The same author's great monograph on "Naval Architecture"[15] contains several valuable chapters on waves. In 1865 there were published the results of experiments made during the preceding decade by Bazin and Darcy[16] on a much more extensive scale than those performed by Russell.

Airy's elaborate treatise "On Tides and Waves"[17] appeared in the Encyclopedia Metropolitana in 1845, and has since been recognized as the standard mathematical discussion of the theory of waves, although the validity of some of his assumptions has been assailed. In spite of its technical character the non-mathematical student will find in it much of value. Two papers by Stokes[18] which appeared a few years later and which have since been included in the first volume of his "Mathematical and Physical Papers," are important because of their contributions to the theory of oscillating waves. Rankine gave a mathematical analysis of the "Exact Form of Waves near the Surface of Deep Water"[19] in 1863. Fourteen years later Boussinesq produced his exhaustive treatise entitled "Essai sur la Théorie des Eaux Courantes"[20] which includes an extended mathematical discussion of waves. Bertin's long "Étude sur la Houle et le Roulis"[21] and still more elaborate "Données Théoriques et Expérimentales sur les Vagues et le Roulis"[22] appeared in sections during the decade 1869-1879 in the Mémoires de la Société Nationale des Sciences Naturelles de Cherbourg, a publication which in the same period carried articles on the same or related subjects by de Saint-Venant[23], Mottez[24] and others. All of these papers except the last mentioned are mathematical in character, but contain matter of importance for the non-mathematical student of wave action, the later sections of Bertin's second memoir including the results of experiments made by himself and Caligny upon the effects of waves breaking on sloping beaches, either with or without the disturbing effects of seawalls.

In 1866 Cialdi published his important book "Sul Moto Ondoso del Mare e su le Correnti di esso"[25], in which he reviews the works of many previous writers, particularly those of Italian

authors, and discusses wave action from the standpoint of the engineer. Caligny's important work on "Oscillations de l'Eau,"[26] published in 1883, includes the results of valuable experimental work on waves, particular interest attaching to his contributions to our knowledge of waves of translation. Stevenson's treatise on "The Design and Construction of Harbours"[27] contains a large number of facts which have materially increased our familiarity with the mechanical work of waves, and from the engineering point of view is one of the best published treatises on wave action. A little book on "Waves and Ripples in Water, Air and Aether"[28] by Fleming, although representing a course of lectures given before a juvenile audience, presents in simple form many laws of wave motion which will interest the older reader. Wheeler's "Practical Manual of Tides and Waves"[29] reviews a few of the important works on waves, and discusses the principles of wave action at some length. A large number of interesting facts concerning the behavior of waves will also be found in the same author's volume on "The Sea Coast"[30]. Vaughan Cornish's beautifully illustrated book entitled "Waves of the Sea and other Water Waves"[31] does not consider the principles of wave motion very fully, but presents a wealth of facts concerning the height, length, and other elements of waves, and discusses the action of waves on shore detritus.

The best general review of the principles of wave action which has come to my notice is to be found in the second volume of Krümmel's "Handbuch der Ozeanographie"[32]. Gaillard's treatise on "Wave Action in Relation to Engineering Structures"[33] contains a fairly extended review of the most important work of previous writers and discusses the results of the author's own excellent researches. The book loses part of its value as a reference work because many of the quotations from the works of previous writers are unaccompanied by such citations of the original sources as would enable the reader to find them. White's "Manual of Naval Architecture"[34] has a valuable chapter on deep sea waves. The numerous papers by Vaughan Cornish, published in the London Geographical Journal and elsewhere, contain many interesting facts not stated in his book above mentioned; and the volumes of the "Proceedings of the Institution of Civil Engineers" (London) include a number of extended articles on the action of waves and currents upon

shore débris, which together with the voluminous discussions appended, present various facts and theories of interest to the student of wave action. Many other sources on which I have drawn are mentioned in the pages which follow.

WAVES OF OSCILLATION

Origin. — The waves produced by the action of the wind are the most important type of sea waves. When wind acts upon a water surface it subjects it to irregular, unequal pressure because winds never blow with constant velocity, but always in irregular gusts. Unequal pressures deform the water surface, giving it an undulatory form. The wind can then act directly upon the undulations, pressing strongly against the sides of the elevations, but acting less effectively against the partially protected depressions. The water in the elevations is moved forward, both by direct pressure and by friction with the passing air. This action causes the undulations to advance and to increase in size until the limit of wave height for the given wind velocity is reached, providing the breadth and depth of the water body are sufficiently great.

If one watches the surface of a pond when a faint breeze first springs up, he will note that the once glassy surface suddenly becomes covered with tiny ripples, which disappear almost as suddenly if the breeze dies down. But if the breeze continues, it will be seen that these miniature waves increase in size progressively toward the leeward side of the pond, those on the windward side remaining the original size. If the breeze now ceases suddenly, the tiny ripples on the windward side quickly vanish, but the larger waves developed where the wind blew across a greater expanse of water continue to agitate the surface of the pond for some time. It can be shown that the wind has produced two distinct types of waves. The tiny ripples belong to the class known as capillary waves, are like the ripples produced by a falling raindrop or a fine wire moved through the water, are due to surface tension rather than to gravity, and move the more rapidly the smaller the wave length. On the other hand, the larger waves on the leeward side of the pond belong to the class usually denominated by the term "waves of oscillation," are due entirely to gravity, move the more rapidly

the greater the wave length, and very large examples in the ocean may travel for hours or days without any sensible loss of energy due to viscosity. There is a certain length of wave, therefore, on the border line between large ripples and small waves of oscillation, which has the slowest rate of motion. Progressively shorter waves travel with increasing velocities and belong to the class of ripples. Those of progressively greater length also travel with increasing velocities, but belong to the class of true waves of oscillation[35]. A good brief summary of the principal points in the theory of oscillatory waves will be found in a paper published by Lyman in 1868[36].

Wave Motion. — In all types of waves, the wave form moves far over the surface of the water while the individual water particles move but a comparatively short distance; just as "waves" may be seen to sweep across a wheat-field with every gust of wind, although the individual stalks of wheat merely bend slightly and then return to their original positions. The contrast between wave movement and water movement is strikingly exhibited when waves advance up an estuary during the ebbing of the tide. In typical waves of oscillation in deep water each water particle moves through a circular orbit, the particle moving forward on the crest of the wave, downward on the back, backward in the trough, and upward on the wave front. The relation of the orbital paths of the water particles to the direction of wave propagation is shown in Figure 1. It is important to note with care both the direction of orbital motion, and the part of the orbit in which a water particle has a given direction, as these points frequently are incorrectly represented. For example, one of our best known college texts on "Physiography" contains a figure illustrating wave motion which erroneously shows the direction of orbital movement at the crest of the wave as opposite to the direction of wave propagation, while the black dots representing the water particles are in the wrong positions in all of the orbits except those showing the particle at the top of wave crest and bottom of wave trough.

A cork or piece of seaweed floating on the water, and moving with the water particles, may be seen to describe a circular orbit when a wave form passes under it. The cork is at the top of its orbit as the crest of a wave passes, reaches the bottom as the trough passes, and attains the top when the next crest

arrives. Thus the time required for the cork to move through its orbit is precisely that required for the crest of the wave to advance a distance equal to one complete wave length, i.e., the distance from the crest of one wave to the crest of the next. Now in a wave 20 feet high, having a length of 1000 feet or more, it is evident that the water particle travels through its circular orbit a distance of but little more than 60 feet while the wave form travels a fifth of a mile. As we shall see in a later paragraph, the velocity of waves is often so great that the ocean would be unnavigable were it not for the fortunate fact that the water does not travel with the wave form.

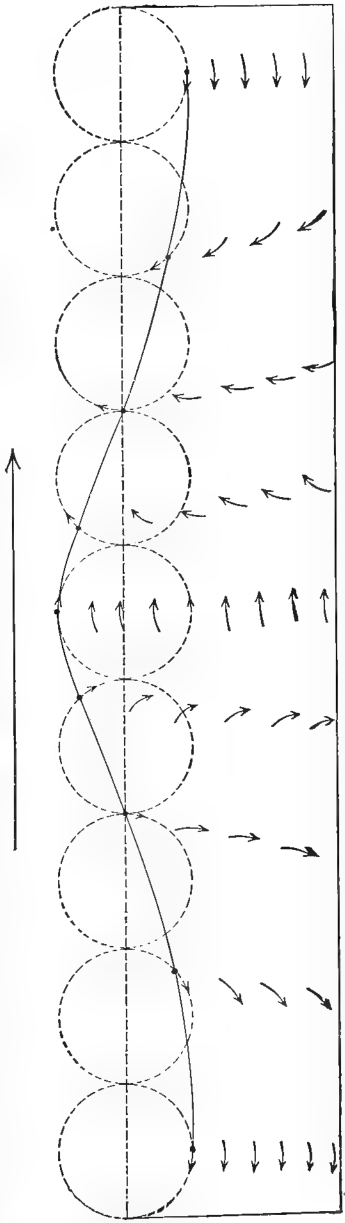

Fig. 1. — Diagram showing the motion of water particles in an oscillatory wave. The large arrow shows the direction of wave advance, the small arrows the direction in which water particles are moving in different parts of the wave.

Although emphasis is properly laid upon the fact that the particles of water move in a limited orbit while the wave form progresses, the common statement that in the open sea the water particles have no progressive motion is not quite accurate. In 1847 Stokes demonstrated from the mathematical standpoint that "the particles, in addition to their motion of oscillation, will have a progressive motion in the direction of propagation of the waves"[37], the forward motion of the

particles being not altogether compensated by their backward motion. According to Stokes this progressive motion, in deep water at least, decreases rapidly as the depth of the particle considered increases. Cialdi later discussed this progressive motion of the water particles at much length, and sought to explain it as in part a consequence of the increase in density of the particles brought about by the cooling due to evaporation and radiation at the crests of the waves[38]. It is certain that the wind by pressing more upon the posterior parts of the waves than upon the anterior parts, gives a distinct progressive motion to the water involved in oscillatory waves, and that this motion is greatest at the surface, decreasing with depth. Stokes has developed a formula for calculating the extent to which a ship may be drifted from her course by the progressive motion of the water particles in waves of this class, although he does not regard the formula as of practical importance[39].

In water of limited depth the water particles move round and round in ellipses whose major axes are horizontal (Fig. 2),

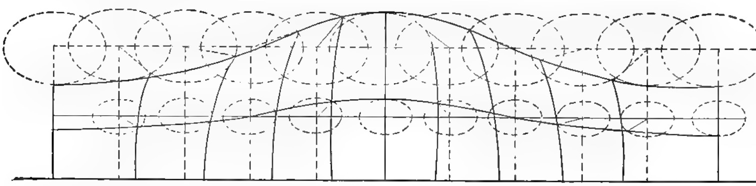

Fig. 2. — Diagram showing the elliptical orbits of water particles in shallow-water waves, and the decrease in size of orbits with increasing depths. (After Krümmel.)

and at the bottom the ellipses are reduced to straight lines, the water particles simply moving forward and backward[40]. In somewhat deeper water the particles near the surface will move in circles, those farther down in ellipses, and those on the bottom in straight lines. It is this back-and-forth movement on the bottom which Emy[41] was considering when he proposed his theory of " groundwaves " or " bottom waves " (flots de fond), although he apparently included in addition certain phenomena of waves of translation. This theory was assigned an undue importance, and was greatly elaborated by Cialdi[42] and Cornaglia[43], and by others of the Italian school whose works discuss the " flutto di fondo " at much length. The latter author lays

much stress on the existence of a "neutral line" where the landward and seaward components of the groundwave are supposed to be exactly balanced; and considered that inside this line the motion of debris is landward, while outside it is seaward. Thoulet[44] applies the term "lames de fond" to waves of an entirely different type, — waves originating from seismic disturbances, the discussion of which will be taken up on a later page. On a level sea-bottom covered by a limited depth of water, it is evident that oscillatory waves would cause sand to shift back and forth, but would give to it no progressive motion, were there no progressive motion of the water particles themselves. If we admit the existence of the progressive motion discussed in the preceding paragraph as characteristic of normal waves of oscillation, it would seem to follow that this motion will still obtain when the orbits are reduced to straight lines, and that we should therefore expect, in the absence of opposing forces, a slow but progressive transfer of sand in the direction of wave advance.

Caligny investigated a series of waves formed by raising and lowering a cylinder in the end of a wooden trough, and found that the water particles moved in elliptical orbits which had their greatest diameters *vertical* instead of horizontal. It is possible that the orbital motion of this type of wave is responsible for those illustrations of sea waves appearing in certain text-books of physical geography, in which the orbital paths are shown as ellipses, with major axes vertical. But according to Caligny[45] these waves are peculiar in several respects: they belong to the class of waves of translation, although they have an oscillatory movement; and experiments showed that grains of sand and other material were slowly transported along the bottom of the trough in a direction *opposite* to that of the wave propagation. It would seem inadmissible to compare these waves with those formed by the wind in the open ocean. Bremontier[46] supposed that in normal wave motion the water particles rose and fell in vertical paths, while Emy[47] presents arguments to show that the paths must be ellipses with the major axes vertical. In both cases the arguments are evidently unsound, and the conclusions opposed by the results of more modern studies of deep-sea waves. In short, I have not found a satisfactory basis for those illustrations of deep-sea waves showing elliptical orbits with major axes vertical.

As will readily appear from Figures 2 and 3 the size of the orbits through which the water particles move decreases rapidly with increase in depth. At the depth of one wave length below the surface, the water particles of an oscillatory wave are moving in orbits whose diameters are only $\frac{1}{534.5}$ as great as the diameter of the orbits at the surface[48]. We may express this relation in

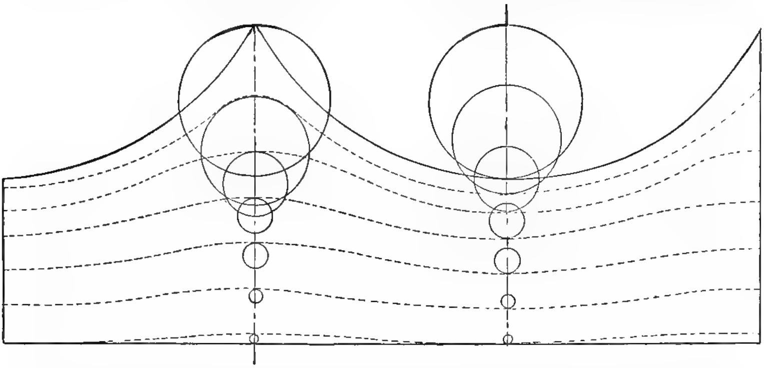

FIG. 3. — Diagram showing theoretical form of a cycloidal wave, and the rapid decrease in size of the orbits (through which the water particles move) with increasing depth.

the following rule[49]: For each additional $\frac{1}{9}$ of the wave length below the mid-height of the surface wave, the diameter of the orbit is decreased by $\frac{1}{2}$. Thus:

Depth below mid-height of surface wave in fractions of wave length	0, $\frac{1}{9}$, $\frac{2}{9}$, $\frac{3}{9}$, $\frac{4}{9}$, etc.
Proportionate diameter of orbit	1, $\frac{1}{2}$, $\frac{1}{4}$, $\frac{1}{8}$, $\frac{1}{16}$, etc.

For the diameter of an orbit situated one wave length below the surface, the rule would give a value of $\frac{1}{512}$ of the surface orbit, which is approximately correct and is the figure quoted by Cornish[50] and others. If the sea is disturbed by waves having a height of 20 feet and a length of 400 feet, the water particles at the surface move in circles having a diameter of 20 feet, while the particles at a depth of 400 feet move in circles only $\frac{4}{10}$ of an inch in diameter. The importance of this principle will appear when we come to consider the depths at which waves may erode the sea-bottom and transport material.

Wave Form. — The theoretical form of oscillatory waves in the open sea is indicated by Figure 4 which represents the profiles

of three such waves. The profiles are *trochoidal* curves[51], or the curves which would be described by points within a circle which is rolled along the under side of a straight line. In the figures this curve is produced by drawing a series of circular orbital

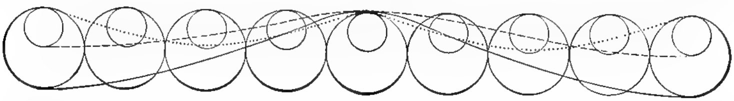

FIG. 4. — Theoretical profiles of three trochoidal waves having different sized orbits (solid-line profile and broken-line profile), or different spacing of orbits (broken-line profile and dotted-line profile). Modified after Grabau.

paths, indicating the proper position of the water particle in each orbit, and connecting these positions by a curved line. As will appear from the figures, the sharpness of the wave crests varies according as the series of orbits having water particles in the same given positions, is closely or widely spaced. From the mathematical standpoint, the curve will be sharp crested or not according as the point within the rolling circle is at or near the circumference, or near its center. If at the circumference, the curve developed will be the very sharp crested form called the *cycloid* (Fig. 3). This is the shortest and steepest form which a true wave theoretically can have[52]. As a matter of fact no wave approximating the form of the common cycloid can be produced in nature, as Gaillard has shown[53]. In the steepest deep-sea waves observed the ratio of height to length is only about one-half that demanded by the cycloidal wave form[54]. It is doubtful whether the precise form of the flatter trochoid is ever achieved, for it can be shown that the trochoidal theory of waves does not adequately satisfy all the conditions of wave formation[55]. Nevertheless, the deviation of deep-water waves from the true trochoidal form is so slight, and the trochoidal theory, especially as modified by Stokes[56], is so superior to all other theories of wave formation, that we shall not go far wrong if we consider such waves as having the form of the trochoid and call them trochoidal waves.

In any trochoidal wave the crest is steeper and narrower than the trough and contains an insufficient amount of water to fill the trough. The level of the water during calm is therefore lower than the level of the centers of the orbits which the

PLATE I.

Photo by Judges.

Storm waves breaking against sea wall at Hastings, England.

surface water particles describe during wave action. In other words, half the height of the waves does not give the true sea level, that level being somewhat lower. Stevenson gives a formula prepared by Rankine, for calculating the position of mean sea level when height and length of wave are known; he also observes that large waves in Wick Bay had about two-thirds of their height above still-water level, and one-third below[57]. On the basis of extensive observations Gaillard has devised more satisfactory formulæ for determining the still-water level, taking due account of the fact that a larger percentage of wave height is above still-water level in shallow water than would be indicated by a formula which, like Stevenson's, is applicable to deep-water waves. Gaillard found that in shallow water about three-quarters of the wave height is above still-water level just before the wave breaks[58]. The importance of this fact will be apparent when it is remembered that the effective salt-water level of the sea may thus be raised a number of feet above high tide level, and also that floating logs or blocks of ice may accomplish considerable work to any height reached by the crest of the waves.

When a strong wind is blowing, the trochoidal profile of the waves is seen to be materially altered. If the wind is in the direction of wave propagation, as is more commonly the case in the open sea, the forward motion of the water particles on the wave crest is accelerated, while the backward motion in the trough is retarded. Since the troughs are somewhat protected from the wind, the retardation is less effective than the acceleration of the wave crests. The net result is a steepening of the front of the wave, so that the profile becomes noticeably asymmetrical. Winds of sufficient velocity may even force some of the water on the wave crest out of its orbital path, blowing it forward into the adjacent trough in the form of foam and spray. When asymmetrical waves pass out of the region of the storm winds which generated them, they decrease in height, become more rounded and symmetrical, and closely approach the trochoidal form, although the steeper front has been observed on deep-water waves in calm weather[59]. These waves may be propagated hundreds or thousands of miles from the storm center where they originated, and ultimately become the gentle undulations known as the "swell," or "ground swell."

Surf. — A very important alteration of form occurs when the oscillatory wave passes into shallow water. The wave becomes higher and shorter, the front steepens, the crest arches forward and, finding itself unsupported by sufficient water on the front of the wave, dashes downward with a roar, producing the phenomenon known as the "surf." An individual breaking wave is known as a "breaker," or less frequently as a "combing wave"; the latter term is also applied to a deep-water wave whose crest is pushed over forward by a strong wind. The commonly accepted explanation of surf is that the wave is retarded by friction when it enters shallow water, the lower part "dragging" on the bottom while the upper part advances unimpeded, until the wave becomes so steep in front that it falls forward. There seem to be fatal objections to this theory of surf action. In the first place the amount of friction necessary to produce the observed result does not seem to exist. Experimental studies of waves in shallow water of uniform depth under conditions favorable for the development of frictional retardation fail to show it[60]. On the other hand, it will later be shown that wave velocity decreases with decreasing depth. It is equally certain that the size of the orbital paths increases as waves enter shallower water, while at the same time the volume of water is decreasing. With constantly enlarging orbits and diminishing water supply, there must come a time when the volume of water is insufficient to build up the entire wave form, the deficiency manifesting itself as a "hollowing" of the front of the wave. The water available endeavors to curve around through the entire orbit, but on reaching the top of the circle finds itself unsupported and collapses.

The form of a breaking wave is not that which should exist if friction were the principal cause of the surf. If the observer can secure a position where the wave profile is discernible, he will find that there is a steepening of the wave front, to be sure, but the form does not suggest a steepening due to forward inclination of the whole wave mass resulting from "bottom drag," so much as it does a steepening due to the absence of water on, and consequent hollowing of the front side of the wave. When the wave finally breaks, masses of foam floating on the water surface appear to describe an orbit that is more symmetrical than one should expect in a wave deformed by great bottom friction, while the forward arching crest tries to complete a

PLATE II.

Combing wave, showing water completing orbital movement although insufficient in quantity to fill the wave form.

wave form which, if achieved, would not show excessive steepening on the front. (Plate II.) The credit for first stating the above explanation of surf action belongs to Hagen[61].

Depth at Which Waves Break.—The depth of water in which the oscillatory wave assumes the form of a breaker is a matter of some interest. As in the case of the wave of translation, described below, Russell[62] found that breaking occurred when the depth of the water equalled the height of the wave, a rule not wholly confirmed by the experiments of Bazin[63], who found that breaking occurred more frequently when the height of the wave exceeded two-thirds of the total depth. Russell states that his rule also holds good for oscillatory waves, but unfortunately he is neither clear nor consistent in his method of calculating wave height and water depth in the case of these waves. In one place we read that "every wave broke exactly when its height above the antecedent hollow was equal to the depth of the water," the method of calculating water depth not being stated; on another page both wave height and water depth are apparently measured from mean water level; according to a third statement the author never saw a wave as much as 10 feet high in 10 feet of water, nor 20 feet high in 20 feet of water, although he has seen waves approach very nearly to those limits[64]. Cornish expresses the rule as follows: waves break when the depth of water reckoned from the undisturbed sealevel is equal to the height of the crest above the trough[65]. In other words, a wave entering shallowing water increases in height as the water decreases in depth until the height of the wave above the trough, and the mean water depth reach approximate equality, when the wave breaks. According to this rule the navigator who sees waves 8 or 9 feet high (or about 6 feet above still-water level) breaking over a certain submarine bar, may know that he can count on but 8 or 9 feet of mean water depth, or 6 feet of depth below the trough, at the place in question. Some other factor or factors, however, combine with water depth to determine the breaking of a wave, with the result that the above rule does not always hold. Departures from the rule are noted by Stevenson[66]. Cialdi[67] cites a great number of cases in which waves have been known to break in water many times deeper than the wave height, and both Thoulet[68] and Krümmel[69] have placed some of these in tabular form. The latter author suggests that the frequent

PLATE III.

Photo by A. M. Cromack.

Waves breaking against seawall at Scarborough, England.

breaking of waves in deep water just above the outer edge of a submarine terrace may be due to an upward push imparted to the lower water when it comes against the terrace face, this push being transmitted to the surface and causing the waves to break[70]. Gaillard found that while oscillatory waves sometimes break quite uniformly when the true height of the wave equals the depth of the water measured from still-water level, in other cases they break when the ratio of water depth to wave height is from 1.16 to 2.71. He observed that the depth at which breaking occurs varies with variations in wind velocity, slope of bottom, smoothness of bottom, and wave length; and suggests that the strength of the undertow is probably another important factor in determining the depth at which waves break. In addition to his own observation Gaillard quotes those of many other observers[71]. The depth of breaking is of importance in determining the position of barrier beaches and other related shore forms.

Intersecting Waves.—Thus far we have considered the form of waves from the standpoint of changes in profile. If now we turn to their variations in form along the crest line, we have first to note that the typical oscillatory wave can not be traced far in the direction indicated. The crest soon descends at either end and is lost in the maze of other waves. In the open sea one experiences the greatest difficulty in determining the end limits of a given crest, and also in following the progressing crest for any length of time. The reason for this is found in the fact that more than one set of waves are always disturbing the ocean surface, and the several sets intersect each other at various angles. Even with two intersecting series it is evident that the water will rise very high where crest coincides with crest, will fall very low where trough coincides with trough, and will have all intermediate elevations where different parts of the front and back of one wave intersect different parts of another wave. Imagine several series of waves crossing each other at distinctly different angles, and we have an adequate explanation for all the great irregularity in wave form observed in the open ocean. Only when the observer is stationed high above the tossing waters, and then only under favorable conditions, can he distinguish the several orderly systems of waves which are responsible for the apparent chaos.

But even in a single wave system the crests are not of indefinite extent. This is because the wind which causes the waves is never of uniform strength, and because the large waves result in part from unequal combinations of smaller waves, as shown on a later page. The wind comes in gusts of varying strength and somewhat varying direction, and so irregular a force could not produce a regular wave crest stretching far over the ocean. Instead we have a large number of short, nearly parallel, overlapping crests which in course of time combine into a smaller number of larger but decidedly irregular waves. Even in the region of the trade winds, where the winds blow with an unusual degree of regularity, "the open sea does not present a series of parallel ridges, each one of uniform height, with a lateral extension many times greater than the distance from crest to crest"[72]. On the contrary, there is no evidence of any continuous approximation toward regularity.

Wave Height. — In discussing the sizes of waves we have to do with two principal elements of wave form: the *height* measured from the bottom of the trough to the top of the crest; and the *length* measured from crest to crest, or from trough to trough. The initial height of the oscillatory waves depends on: (1) the strength of the wind, (2) its duration, and (3) the extent of open water over which it blows. A faint breeze sets in motion very small waves which increase in size to a certain limit, but which would never become great billows. In the trade wind belt the maximum height of wave for a certain strength of wind is soon reached, and although the wind may continue steadily for days at the given strength, there is no increase in the size of the waves. In a general way, the velocity of the wind in statute miles per hour divided by 2.05 will give the height of the waves in feet[73]. Thus the average height of waves in a gale blowing 44 statute miles per hour is

$$44 \div 2.05 = 21.5 \text{ feet.}$$

It should be noted, however, that in very severe storms the highest waves may not occur when the wind velocity is at a maximum, but are seen to develop as the wind begins to subside. The explanation of this phenomenon is probably to be found in the fact that the excessive force of a violent wind blows off the tops of the waves and casts them into the preceding troughs,

thereby materially diminishing the wave height. It is possible also that as the storm subsides the waves, which were compelled to remain independent and irregular under the gusty force of the storm wind, gradually combine into a smaller number of larger waves which are little affected by the failing strength of the dying wind[74].

Effect of Wind Duration.— Wind duration is another factor in increasing wave height up to the limiting height for a given wind strength. When a breeze springs up, small ripples first appear over the water surface, but gradually develop to larger size without any increase in the strength of the breeze. If a large swell is already running in the direction of the wind, a sudden increase in wind velocity results in increased height of waves; but in this case the wind does not have to endure very long to bring about a very remarkable increase in height. Cornish has recorded an increase of 7 feet in the height of waves during a squall lasting 4 minutes, and an increase of 2 feet per minute in the height of waves during another squall[75]. The precise method by which small wind waves grow to large ones is not wholly understood, but the Weber brothers give the following four causes for wave enlargement: (1) the continuous horizontal pressure of the wind upon the wave crest, thus tending to enlarge the orbital movement of the water particles; (2) the combining of several smaller waves moving in the same direction; (3) the pressure exerted by a large wave upon the next following smaller wave, by which the latter is enlarged; and (4) the crossing of waves proceeding in different directions[76]. Cornish thus states another theory of wave enlargement: "The horizontal velocity of the air being greatest at the crest, the downward pressure of the atmosphere is least there. Conversely at the trough, where horizontal velocity is least, downward pressure is greatest. Hence the trough is pushed farther down and the crest is sucked up"[77].

Effect of Length of Fetch.— Of corresponding importance is the effect of "length of fetch" of the wind across open water upon wave height. We have already seen that when a breeze blows across a pond there first appear small ripples over all its surface but that these soon increase in size progressively toward the leeward side of the pond. The ripples on the windward side, where the wind has blown across a small expanse of water only, remain small no matter how long or how strong the breeze may

blow. But those on the leeward side, where the fetch of the wind across open water is greater, soon develop into waves of some size because here the waves due to the direct effect of the wind are combined with the waves originating on the opposite side of the pond and propagated by gravity in the direction of the wind. This illustrates on a small scale a matter of much importance in the case of sea waves. Stevenson has shown that for ordinary gales and distances the height of the waves in feet is 1.5 times the square root of the distance in nautical miles which the wind has blown over open water[78]; or

$$\text{height} = 1.5\sqrt{\text{distance}}.$$

Gaillard observed waves 23 feet high near Duluth with a length of fetch of 259 nautical miles[79]. This agrees fairly well with the calculated height of 24.1 feet based on the formula. Des Bois prepared a table to show the heights of waves corresponding to different wind velocities, based on his observation that a wave 2 meters high corresponded to a wind velocity of 5 meters per second, and the provisional theory that " the square of the velocity of the wind will be proportional to the cube of the height of the wave"; and he found that this table corresponded roughly with the results he obtained from a large number of direct measurements[80].

For short distances a modification of Stevenson's formula is necessary. The following table is condensed from one given by that author, and shows the appproximate heights of waves as determined by length of fetch, assuming great depth of water and a strong gale of wind.

TABLE SHOWING APPROXIMATE HEIGHTS OF WAVES DUE TO DIFFERENT LENGTHS OF FETCH

Nautical miles	Heights in feet	Nautical miles	Heights in feet	Nautical miles	Heights in feet
1	3	5	4.3	50	10.6
2	3.4	10	5.6	100	15
3	3.8	20	7.1	200	21.2
4	4.1	30	8.4	300	26
		40	9.5		

For expanses of open water exceeding 500 or 600 miles in length the height of storm waves does not appear to increase

according to Stevenson's empirical formula. With a fetch of 3600 miles the waves should reach a height of 90 feet, but so great a height is probably never attained. The reason for this discrepancy is doubtless to be found in the fact that we have no storm winds blowing steadily for a long period in the same direction over so great a stretch of water[81]. The facts that the wind direction may be approximately the same over a long stretch of water, or that it may have a constant direction for several days at a given place, as noted by Redfield and by Stevenson[82], are not alone sufficient. The winds must blow with the strength of a strong gale in a constant direction over the entire distance for several days, if the full effect of a 2000 or 3000 mile "fetch" is to be realized, since the waves formed to windward must have time to travel the long distance to leeward and produce the cumulative effect which results in maximum wave height. In our cyclonic storms the greatest distance traversed by heavy winds in a reasonably constant direction and for a period of time sufficient for large waves to develop, probably does not exceed 600 or 700 miles. The "effective fetch," therefore, is much more limited than the absolute distance across open water; and Vaughan Cornish has estimated, from a study of charts illustrating weather conditions in the North Atlantic Ocean for nine weeks of exceptionally stormy weather, that the greatest effective length of fetch during that period was about 600 nautical miles[83]. But while waves formed on greater expanses of open water do not reach the heights calculated from the formula given above, they do exceed the altitude of about 37 feet calculated for the greatest effective fetch, because they may combine with an already existing swell and thereby increase their height.

Recorded Wave Heights. — Observations of the heights of waves are often unreliable, but the approximate height under different conditions has been pretty well established by a number of competent observers. On Lake Superior, waves reach a height of from 20 to 25 feet[84]; in the Mediterranean Sea, 25 to 30 feet[85]. Scoresby's oft-quoted observations on the North Atlantic give a height of 43 feet for the largest waves[86], and Cornish reports waves 43 feet high from the same ocean[87]. When two great waves intersect, peaks of water may rise momentarily to a height of 50 or even 60 feet[88]. Although the North Pacific Ocean has a breadth of open deep sea

much greater than that of the North Atlantic, the waves do not appear to reach any greater height[89]; but in the Southern Ocean waves attain heights of from 45 to 50 feet[90]. White refers to trustworthy observations of waves of a single series having heights of 44 to 48 feet, and mentions waves formed by the combination of two or more series said to attain from 58 to 65 feet[91]. Gaillard gives an interesting tabulation of the height, length, and period of ocean waves recorded by a number of different observers, the highest figure for wave height in the table being "greater than 50 feet," in the case of a wave photographed by Capt. Z. L. Tanner of the U. S. Navy[92]. By means of a barometer Abercromby measured waves 46 feet high in the Southern Ocean, and concluded that some waves certainly attain a height of 60 feet[93]. Airy was of the opinion that under no circumstances does the height of an unbroken wave exceed 30 or 40 feet[94]; but against this theoretical opinion we may safely accept the figures of competent observers, and conclude that waves 40 feet high are of fairly frequent occurrence in the open ocean, while heights of 50 feet or more are rare, but not unknown.

When these high storm waves run out of the storm area, they gradually decrease in altitude, and in the form of swells usually do not exceed a height of 15 or 20 feet. By the time they are nearing a distant coast they may have been reduced to heights of a few feet only, and so have become almost imperceptible. Entering shallowing water they seem to awaken to new life, crowding closer together and increasing in height until they break. At the time of breaking the wave height may be anywhere from a few feet up to 25 feet or more. If a wave comes in contact with a vertical wall or cliff the base of which reaches down to deep water, the wave is reflected back without breaking. The water next the wall moves up and down through a vertical distance equal to twice the original height of the wave, as does also the water half a wave length from the wall. Similarly, a wave running in a direction parallel to a vertical or steep wall has that portion of the wave next the wall notably increased in height[95].

Combined Waves. — Waves which appear to belong to the same series vary greatly in height. The larger figures given above are for individual waves, and in each case the average height for the series to which the waves belonged was much less. Thus a wave 40 feet high may occur in a series of waves having an average height of

but 20 or 25 feet[96]. This inequality in wave height is probably due in considerable part to the fact that what appears to be a single series of waves of irregular height is really the combined effect of two or more series of waves moving in the same direction, each series having different but fairly constant height and length. Figure 5 from Cornish's work on "Waves" shows, in the third line, the profile of an apparently irregular series of waves (*c*) resulting from the combination of the two regular series (*a* and *b*) shown in the first and second lines. By holding the page with the figure nearly on a level with the eye, but slightly inclined toward the observer, the marked irregularity of the combined series may easily be detected.

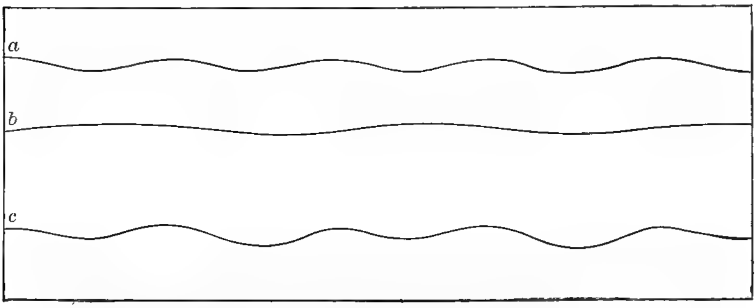

FIG. 5. — Diagram showing how two regular series of waves (*a* and *b*) of different heights and lengths combine to form an irregular series (*c*).

The successive wave heights, in feet, measuring from each crest to the next trough, toward the right, are as follows: 22.50, 37.50, 18.75, 10.00, 27.50. The average wave height for an indefinite length of this irregular series will be 30 feet or precisely the height of the dominant regular series. It is evident, therefore, that an observer might conclude that the sea was disturbed by a single series of waves 600 feet long and 30 feet in average height, and that the real presence of a swell 20 feet high would be undetected. This may explain the fact that during a storm at sea the long swell remains invisible, yet becomes noticeable as soon as the shorter storm waves die down a little[97]. Gaillard suggests, however, that the waves first made by a strong wind are of unstable form and cannot travel far without being destroyed and contributing their energy to the more stable waves of nearly perfect trochoidal form, the "swell"; while Taylor is of the opinion that direct wind action causes the

water particles to move in orbits of varying amplitude and velocity, producing a confused sea; but that as soon as the wind ceases the viscosity of the water tends to make the orbits identical, and thus to produce a more uniform system of trochoidal waves[98].

The combination of two or more series of waves moving in the same direction explains the fact that when waves break upon the shore, there is a recurrence, at intervals, of waves of exceptional height. It should be noted, however, that while the popular idea that every seventh wave is a big one rests upon a basis of fact, the ratio of wave lengths in the combining series is just as likely to make every third, or ninth, or some other wave the largest; or if three sets of waves combine, the large waves may arrive at irregular intervals.

Wave Length. — The total energy of a wave has been shown to vary nearly as the square of the height and as the first power of the length, so that these dimensions may be said to measure the capacity of a wave for destruction[99]. Of these two important elements of wave form we have just considered the height, and may now turn our attention to the length.

The ratio of wave height to wave length is a matter of considerable interest. Inasmuch as storm waves usually appear higher and steeper than those in a moderate sea, we should expect this ratio to increase with increasing roughness of the sea. Lieutenant Paris found that in a light sea the ratio of height to length is only 1 to 39, in a rough sea 1 to 21, while in a heavy sea it rises to 1 to 19[100]. Schott compared the ratios directly with the strength of the wind, and found that with a moderate wind the ratio of height to length was 1 to 33, with a strong wind 1 to 18, and with a storm wind 1 to 17 or even as high as 1 to 13[101]. On the other hand, White compares the ratios with the lengths of the waves, and shows that as the lengths increase the ratios diminish. Thus he finds from an analysis of 179 published French observations that with a wave length of less than 100 feet, the average ratio of height to length is 1 to 17; with a length of 100–200 feet, the ratio is 1 to 20; with a length of 200–300 feet, 1 to 25; with a length of 300–400 feet, 1 to 27. For greater wave lengths the figures are not wholly in accord with the theory, while in waves from 100 to 400 feet long the very small ratio of 1 to 50 has been observed[102]. Cornish has com-

pared the lengths of waves with the expanse of open water over which the wind blows and finds that " the length of the storm-waves is increased when the length of the sheet of water is increased, but more slowly "[103].

The lengths of deep-water waves are quite definitely related to their velocities and to their periods, as will be shown more fully on a later page; but we may note here that the wave length (in feet) is roughly equal to $5\frac{1}{8}$ times the square of the period (in seconds). Thus, if waves pass a given point at the rate of one in every 4 seconds, the wave length must be approximately 82 feet; for

$$\begin{aligned} \text{Length} &= 5\tfrac{1}{8}\ (\text{period})^2 \\ &= 5\tfrac{1}{8}\ (4)^2 \\ &= 82 \text{ feet.} \end{aligned}$$

Recorded Wave Lengths.—The greatest trustworthy measurement of wave length is that recorded by Capt. Mottez of the French Navy, for a wave in the North Atlantic, measuring 2750 feet from crest to crest. In the English Channel Cornish observed waves whose period indicated a length of 2594 feet[104]. Ross observed a wave in the South Atlantic 1920 feet long[105]. The greatest length reported by Des Bois is 1640 feet[106], while Major Leonard Darwin found the waves of an exceptionally severe storm in the Southern Ocean to be 1200 feet in length[107]. Some of these high figures are probably due to the combination of two sets of waves in such a manner as to give an abnormally long stretch of low water between two crests, for storm waves in the open sea are not usually more than 600, and very rarely more than 700 feet long. Scoresby found the extreme length of the great storm waves measured by him to be 790 feet[108]. Officers on the North Atlantic liners regard 600 feet as an enormous wave length, although they agree that larger lengths are to be found in the Southern Ocean, where in one exceptional storm Lieutenant Paris found the greatest average length was 771 feet, with not a few waves over 900 feet, and several surpassing 1312 feet in length[109].

There seems to be little doubt, however, that the swell has a length often more than double that of storm waves, and at least one of the figures given above, that of 2594 feet for the length of waves observed by Cornish, refers to the swell. When the swell enters shallow water the velocity and wave length are

diminished, but the period remains the same. Since the period bears a definite relation to the length of the waves in deep water, it is possible, by counting the number of breakers arriving at the shore in a given time, to determine the lengths of the waves in the open sea. In this manner it has been established that the swell in the open sea must not infrequently have lengths of from 1000 to 2000 feet, and occasionally more[110]. Now in deep-water waves a great wave length means a great velocity, and some authorities doubt whether short storm waves will lengthen to form the longer swells, since this would mean that the speed of the waves was accelerated after the wind ceased to act upon them. Antoine[111], however, believes that just such an acceleration does occur. Others suppose that the waves are propagated by gravity at the same rate of speed given them by the wind, or even that their velocity suffers a slight diminution. Cornish concludes that the longer swells are present during storms, but are obscured by the shorter waves which are then more prominent[112]. We shall find later that the longer waves, while they agitate the surface less than storm waves, agitate the deeper waters much more, and have an important effect upon the shoreline.

Wave Velocity. — The velocity of oscillatory waves is a matter of considerable interest in various connections. We have already observed that the wave form travels at a speed very much greater than that of the water particles themselves. Thus, a wave 400 feet long and 15 feet high will have a velocity of about 45 feet per second, while the surface water particles will move round in their orbits at a speed of but $5\frac{1}{3}$ feet per second. For ocean waves of large size the wave velocity is apt to be six or seven times as great as the orbital velocity; but it is impossible to give any definite rule for the relations of these two elements of wave motion[113].

We can correlate the velocity of wave motion with wave length more precisely, however, for in deep water the velocity of the wave depends on its length, and is proportional to the square root of its length[114]. The velocity of any wave whose length is known may be calculated approximately by very simple formulæ. Thus, the velocity in miles per hour is equal to the square root of $2\frac{1}{4}$ times the wave length measured in feet[115]. If it is desired to have the result expressed in feet per second, then

the velocity in feet per second is equal to $2\frac{1}{4}$ times the square root of the length in feet[116]. According to the first formula a wave 100 feet long will have a velocity of 15 miles per hour; for

$$\begin{aligned} \text{Velocity} &= \sqrt{2\tfrac{1}{4} \times \text{length}} \\ &= \sqrt{2\tfrac{1}{4} \times 100} = \sqrt{225} \\ &= 15 \text{ miles per hour.} \end{aligned}$$

According to the second formula the same wave will have a velocity of 22.5 feet per second; for

$$\begin{aligned} \text{Velocity} &= 2\tfrac{1}{4}\sqrt{\text{length}} \\ &= 2\tfrac{1}{4}\sqrt{100} = 2\tfrac{1}{4} \times 10 \\ &= 22.5 \text{ feet per second.} \end{aligned}$$

If we reduce the 15 miles per hour, derived from the first formula, to feet per second, we get 22 feet per second, which agrees fairly well with the result obtained by the second formula. We may also determine the approximate velocity of a wave in feet per second by the formula:

$$\text{Velocity} = \sqrt{5\tfrac{1}{8} \times \text{length}}$$

which becomes, in the case of the wave described above,

$$\begin{aligned} \text{Velocity} &= \sqrt{5\tfrac{1}{8} \times 100} \\ &= 22\tfrac{2}{3} \text{ feet per second.} \end{aligned}$$

As Gaillard has pointed out in commenting on the above formula, the velocity of a deep-water wave is practically the same as that which a body would acquire in falling through a distance equal to 8 per cent of the wave length[117].

Because of the relations existing between wave velocity, wave length, and the period of the waves, we may determine the velocity of waves in other ways. Thus the velocity of the wave in knots per hour is roughly equal to three times the period (in seconds)[118]. Or if we transform the period of the wave into the number of waves per minute (wave-frequency), then the velocity in feet per minute is equal to the wave length multiplied by the frequency. The velocity in miles per hour may be found by dividing the frequency into 198[119]. Thus if the wave 100 feet

in length, considered above, have a period of about $4\frac{1}{2}$ seconds, then the velocity in knots per hour is roughly $13\frac{1}{2}$, for

$$\begin{aligned}\text{Velocity} &= 3 \times \text{period}\\ &= 3 \times 4\tfrac{1}{2}\\ &= 13\tfrac{1}{2}\text{ knots per hour.}\end{aligned}$$

The velocity of this same wave in feet per minute will be 1333; for a period of $4\frac{1}{2}$ seconds means a frequency of $13\frac{1}{3}$ ($60 \div 4\frac{1}{2} = 13\frac{1}{3}$), whence we have the following:

$$\begin{aligned}\text{Velocity} &= \text{wave length} \times \text{frequency}\\ &= 100 \times 13\tfrac{1}{3}\\ &= 1333\text{ feet per minute.}\end{aligned}$$

This agrees roughly with the velocities previously obtained, since it is equivalent to a speed of 22.2 feet per second. The velocity as determined from the frequency alone is 14.85 miles per hour; for

$$\begin{aligned}\text{Velocity} &= 198 \div \text{frequency}\\ &= 198 \div 13\tfrac{1}{3}\\ &= 14.85\text{ miles per hour.}\end{aligned}$$

In order to determine the velocity of a set of waves by this last method it is only necessary to count the number of times per minute some floating object bobs up and down as the waves pass under it, or to count the waves as they rise against some fixed object. The result is in sufficiently close agreement with the velocity of 15 miles per hour determined by a preceding formula.

The periods of waves are more easily determined than are length or velocity, for which reason it is convenient to have in tabular form the lengths and velocities of deep-water waves corresponding to given periods. The table on the following page, taken from White's "Naval Architecture"[120], covers all waves of ordinary size.

Velocities of Shallow-water Waves. — In the preceding pages we have discussed the laws controlling the velocities of deep-water waves. Shallow-water waves, or waves whose lengths are great compared to the depth of the water, obey different laws. It is a well-known fact that such waves move less rapidly than deep-water waves, and Gaillard has expressed in tabular form the relative velocities of the two types, assuming equal wave lengths, but varying depths of water for the shallow-water wave, with a minimum depth equal to .05 of the wave length[121]. The velocities

LENGTH AND VELOCITY OF DEEP-WATER WAVES

(After White.)

Period, seconds	Length, feet	Speed of advance	
		Feet per second	Knots per hour
1	5.12	5.12	3.03
2	20.49	10.24	6.07
3	46.11	15.37	9.10
4	81.97	20.49	12.14
5	128.08	25.62	15.17
6	184.44	30.74	18.21
7	251.04	35.86	21.24
8	327.89	40.99	24.28
9	414.99	46.11	27.31
10	512.33	51.23	30.35
11	619.92	56.36	33.38
12	737.76	61.48	36.42
13	865.84	66.60	39.45
14	1004.17	71.73	42.49
15	1152.74	76.85	45.52
16	1311.56	81.97	48.56

of shallow-water waves of this type must be calculated by means of a formula less simple than those given for deep-water waves, since the formula must be applicable to varying depths of water Such a formula, and numerous comparisons of the observed velocities of shallow-water waves with the velocities computed by the formula, are given in Gaillard's treatise on "Wave Action"[122]. When the wave length is more than 1000 times the depth of the water, the velocity depends wholly upon the depth according to Airy, and is proportional to the square root of the depth. The velocity of such a wave is the same as the velocity which a body would acquire by falling through a distance equal to half the depth of the water[123]. This is the law for the velocity of the wave of translation as determined by Russell[124]; and it should be noted that Airy is inclined to regard the wave of translation as merely a variety of the wave of oscillation[125]. It is also interesting to note that while this law is called Airy's law or formula by some, and is named for Russell by others, it was really applied by Lagrange to water waves at least as early as 1788[126], and is therefore better known as the Lagrange Formula. The law does not hold good for very shallow depths, according to Caligny[127]; nor in moving water, according to Möller[128].

The waves generated in the ocean by earthquakes and submarine volcanic explosions have lengths which are great in comparison to the depth of the ocean, and must therefore obey the laws controlling the movements of shallow-water waves. If we determine the velocity of such a wave, therefore, we should be able to secure some idea of the depth of the ocean it traverses. This was first done by Bache, who estimated the mean depth of the North Pacific Ocean (4200 to 4500 meters) from the velocity of a wave produced by the Simoda earthquake in 1854; and later others followed his example in the cases of the Iquique earthquake and the Krakatoa explosion[129]. The calculations are necessarily inaccurate for various reasons, but are nevertheless of considerable interest.

WAVES OF TRANSLATION

Thus far we have confined our attention to waves of oscillation, in which the water particles move forward on the crest and backward in the trough. There is another type of wave which is also of great interest to the student of shorelines, although its importance is not always appreciated. This is the "wave of translation," in which the water particles move forward as the wave passes, but do not exhibit a compensating backward motion. While not important on the open sea, this type of wave is extensively developed in the shallow waters along all coasts, the waves of oscillation generated in deep water frequently becoming more or less completely transformed into waves of translation as they approach the shore.

Form. — The wave of translation was discovered by Russell, and described at length by him in his reports to the British Association[130]. He showed that when a volume of water was suddenly added to the still water in a canal, or when a portion of the canal water was displaced by suddenly plunging a solid body into it, or when the canal water was pushed into a mound by the shoving motion of a boat or of a plate held vertically, a single prominent wave rolled forward over the canal surface. The entire form of this wave rose above the still-water surface of the canal, and included no trough such as constitutes part of the wave of oscillation. A careful examination of the newly discovered wave showed that it differed widely from oscillatory waves in other

respects, and that the motion of its water particles made the name "wave of translation" appropriate. Let us consider briefly the essential characters of this wave, turning our attention first to the nature of the movements executed by the water particles.

Motion. — Immediately before and immediately after the passing of a wave of translation, the surface of the water and the water particles in depth may be quite still. During the passage of the wave the surface water particles rise and move forward, descending again to the original level, but to an advanced position horizontally, where they come to rest. Thus in Figure 6 the particle *a* rises, moves forward and descends to the position *b*. Water particles below the surface move forward the same distance, but their vertical rise diminishes with increase in depth.

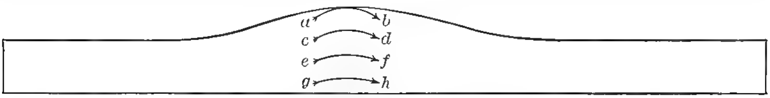

FIG. 6. — Diagram showing movement of water particles in a wave of translation. (After Russell.)

The paths described by the water particles are semi-ellipses which have their major axes horizontal and equal, and their minor axes progressively shorter as the distance below the surface increases, until on the bottom the path becomes a straight line[131]. It will be seen from the figure that water particles vertically above each other, as *aceg*, come to rest in the same relative position farther on at *bdfh*. There is thus a real and permanent forward *translation* of the water itself through a short distance, in addition to the forward *transmission* of the wave form through a very great distance. The space through which the water particles are moved forward is just large enough to contain the volume of water in the wave above still-water level.

Manifestly there are several points connected with the motion of the water particles in waves of translation which will prove of importance when we come to discuss the effect of waves upon shores. The fact that the water particles advance, but do not have a compensating backward motion should result in effective transportation of sand and gravel on shallow sea-bottoms in the direction of wave propagation, unless other forces prevent. It

is likewise worthy of note that in waves of translation the bottom particles move forward just as far as do the surface particles, whereas we have already seen that in oscillatory waves the movement of the water particles dies out rapidly below the surface. Evidently waves of translation may profoundly affect the bottom to great depths, although this conclusion is subject to the qualification, subsequently to be discussed, that waves of translation traversing water bodies of great depth as compared to the size of the waves, tend to be transformed into waves of oscillation. We shall see later that only one-half of the energy of an oscillatory wave is transmitted forward with the wave form, whereas the total energy of a wave of translation is thus transmitted.

Wave Length. - The length of the typical wave of translation is measured from the point where it begins to rise from the still-water level in front to the point where the back slope of the wave again merges with still-water level. These points are not easily determined with accuracy, but according to Russell the wave length thus measured on artificial waves is "equal to about six times the depth of the fluid below the plane of repose." The height of the wave above the still-water surface may be equal to the depth of the fluid in repose, but cannot exceed this measure, as the wave breaks whenever the height becomes equal to the depth[132]. Actual measurements of wave heights and lengths in nature are usually made upon the open sea or in other localities favorable to the formation of waves of oscillation; and while it is possible that some of the figures previously given are really those for waves of translation, no distinction is usually made by the observer, and we lack proper data for the range in size of natural waves of translation.

Velocity. — The velocity of the wave of translation depends upon the depth of the water measured from the crest of the wave, and varies as the square root of the depth. Otherwise expressed, the velocity of the wave is the same as the velocity which a heavy body will acquire by falling freely through a distance equal to half the depth of the fluid below the wave crest[133]. In the deep ocean such waves should have very high velocities, and doubtless many of the earthquake waves which traverse the ocean with velocities from a few hundred miles to nearly a thousand miles an hour[134] are true waves of translation; while the tidal wave which has a velocity of from 480 to 660 miles an hour in depths of

between 12,000 and 20,000 feet[135], is a compound wave having some of the characteristics of the wave of translation. In very shallow water the velocities of waves of translation must of necessity be very low.

Complexities of Waves of Translation.—According to Caligny the waves of translation are not always so simple in character as supposed by Russell. The French engineer made a series of experiments which led him to conclude that there are waves of translation in which the water particles describe closed orbits; that these orbits may approximate vertical ellipses, but that the backward movement of the water particles may slightly exceed the forward movement, causing material on the bottom to be transported in a direction opposite to that of wave propagation; and that a solitary wave of translation may pass through a series of oscillatory waves, complicating their form and causing them to break. He also pointed out that it is possible to have a succession of solitary waves of translation which will resemble ordinary waves of oscillation, the spaces between the waves resembling troughs so closely as to mislead the observer[136].

The investigation of waves of translation in nature is further complicated by the fact that notwithstanding Hunt's arguments to the contrary[137], normal waves of oscillation appear to be gradually transformed into waves of translation when they enter water which slowly decreases in depth, and hence all intermediate phases between the two types of waves may be encountered. When the tops of breakers fall forward, the volume of water thus added to the water surface in front produces waves of translation which run on shore, mingling with waves of oscillation. If waves of translation encounter a cliff or steep shore, they may be reflected, the direction of the transport of water particles in the reflected wave being seaward. For these and other reasons which will presently appear, it may be practically impossible to determine the nature of the water movements which are affecting the distribution of sand and gravel along a shelving coast.

Such complications should not, however, make us lose sight of the importance of waves of translation as agents of shoreline changes. Under favorable conditions the operation of these waves may easily be observed. Thus, when large swells encounter the seaward margin of a submarine terrace (Fig. 7),

they break and form smaller waves of translation which, on a calm day, may cross the shallow water to the shore without deformation until they break as a secondary surf on the beach. The level water surface between any two waves of translation may be seen to differ distinctly from the true trough of the oscillatory wave in deeper water. Russell observed a striking example of waves of translation, formed in the manner above described, on the shore of Dublin Bay, and thus describes the phenomena:

"One of the common sea waves, being of the second order (waves of oscillation), approaches the shore, consisting as usual of a negative or hollow part, and of a positive part elevated above the level; . . . . At length the wave breaks, and the positive part of the wave falls forward into the negative part, filling up the hollow . . . . After a wave has first been made to break on the shore, it does not cease to travel, but if the slope be gentle, and the beach shallow and very extended (as it sometimes is for a mile inwards from the breaking point, if the waves be large), the whole inner portion of the beach is covered with positive waves of the first order (waves of translation), from among which all waves of the second order have disappeared. This accounts for the phenomenon of breakers transporting shingle and wreck, and other substances shorewards after a certain point." Then referring more particularly to the conditions at Dublin Bay, he says that the "waves coming in from the deep sea are first broken when they approach the shallow beach in the usual way; they give off residuary waves, which are positive (waves of translation); these are wide asunder from each other, are wholly positive (i.e., above still-water level), and the spaces between them,

FIG. 7 — Diagram showing how oscillatory waves breaking on a subaqueous terrace produce waves of translation. (After Russell.)

several times greater than the amplitude of the wave, are perfectly flat; and in this condition they extend over wide areas and travel to great distances "[138].

EARTHQUAKE AND EXPLOSION WAVES

In investigations of shoreline changes the student may have occasion to refer to another class of waves which are occasionally developed upon the ocean, and which are improperly called "tidal waves." These are the waves of enormous size and destructive energy produced by submarine earthquakes and volcanic explosions, and for which Hobbs[139] has suggested adopting the Japanese name "tsunamis." They occur at such rare intervals, and operate for such a brief period, that they are probably not of great importance in modeling the forms of the shore. But inasmuch as they temporarily raise the upper limit of salt water far above its normal position, and leave behind them records which may be mistaken as evidences of a former higher stand of the mean sealevel, it is important that we become familiar with the work of these waves.

Nature and Origin of Wave Motion. — A submarine earthquake may produce several types of waves. There are first the short and quick oscillations which travel toward the surface with the velocity of sound in water, and which are felt by overlying vessels as a sharp and violent shock, often causing the sailors to believe that the vessel has struck a reef. Old charts contain many isolated shallows and reefs reported by vessels which had really experienced earthquake shocks in deep water. Occasionally such shocks are severe enough to hurl the ship out of water, to break off its masts, or even to destroy the vessel entirely[140]. These oscillations are not of the type which produce prominent surface waves, however. Other groups of waves are produced by the dislocation of the sea-bottom. While the mechanism of these dislocation waves is not well understood, it is probable that the uplifting of a portion of the sea-bottom raises a mound of water above the general surface of the sea, and that the settling back of this water generates a great wave of translation which traverses the ocean with high velocity. Sometimes several such waves are produced, possibly by the disintegration of a former single wave of translation after the manner described by Russell for some of his experi-

ments[141]. The sudden settling of a submarine crust block may generate a negative wave of translation. On the other hand, the behavior of many earthquake waves upon reaching the coast suggests that they partake of the characters of oscillatory waves, the water particles moving backward in a sort of great trough toward the oncoming wave crest. According to Reid the waves caused by the same earthquake first appear as a depression of the water at some ports, and as an elevation at others; a fact which he attempts to explain on the theory that the down-dropped block generates a negative wave and the upraised block a positive wave[142]. It is possible that the phenomena in question may be explained as a result of the different velocities with which positive and negative waves are propagated, both having resulted from the return of a mass of water raised above the general level, in some such manner as that described by Russell for his "residuary negative waves"[143]. Our knowledge of earthquake waves is still too meager, however, to enable us to speak with assurance on this and other questions concerning their behavior. An experimental study of their mode of propagation will be found in the Weber brothers' "Wellenlehre"[144], a full résumé of our present knowledge of the subject in Krümmel's "Ozeanographie"[145], and a good brief statement in Thoulet's "Océanographie Dynamique"[146].

In submarine volcanic explosions there is also produced a sharp and powerful shock, corresponding exactly to the first mentioned effect of earthquakes. At this time small jets of water may be shot into the air; but there soon follows a doming or up-swelling of the ocean surface, and finally the whole mass of up-raised water may be hurled into the air by the escaping gases. The doming of the water, the push exerted by the gases, and the back-falling mass of water, all tend to produce waves, some of which are waves of translation, and some probably oscillatory or compound waves[147]. Explosion waves and dislocation waves cannot be distinguished, and the origin of many of these waves, often designated collectively as "earthquake waves," remains in doubt. According to Krümmel, Rudolph supposed that the great wave which overwhelmed Lisbon following the earthquake of 1755 was due to a volcanic explosion near the Portuguese coast[148]. Most authorities agree that the waves which followed the eruption of Krakatoa in 1883 were due directly to

the force of the explosion itself, but some have argued that they resulted from the masses of rock falling back into the water[149].

On the open sea the heights of earthquake and explosion waves quickly diminish, and since the lengths are very great, they soon become so low and flat as to be unnoticed by vessels. But when they enter shallow water they behave like other waves, the height increasing until the wave form breaks to produce a gigantic surf. The velocity of these waves is very great, as they may travel a distance of 9000 or 10,000 miles in 24 hours, and one instance is recorded in which a velocity of 900 miles an hour was attained[150]. Their periods range from 15 minutes to one or two hours, and by assuming them to be the periods of free waves in deep water it has been calculated that the lengths of earthquake and dislocation waves vary from 100 miles to 600 miles or more[151].

Recorded Heights. — As students of shoreline phenomena we are more interested in the height attained by this class of waves when they reach the coast. We can better appreciate the truly surprising elevations at which they may leave evidences of their former presence if we review some of the actual cases of which we have authentic records. In the years 358 and 365 A.D., the eastern shore of the Mediterranean was visited by great waves which passed over islands and low shores, sweeping away buildings and thousands of people. Boats were left on the roofs of houses in Alexandria, and others were stranded nearly a mile inland near Modhoni where they were later found slowly decaying[152]. Following the Lisbon earthquake in 1755 a wave variously estimated as from 40 to 60 feet high broke on the coast at Cadiz. The great earthquake at Lima in 1724 was followed by a wave said to have been 80 feet high and which carried four vessels far inland. In August, 1868, an earthquake on the coast of Peru resulted in large waves, one of which submerged the mainland 55 feet above high-water mark. A United States war vessel was carried a quarter of a mile inland at Arica, where it remained until another great wave carried it still farther inland in 1877. This last was the wave caused by the Iquique earthquake, and it is said to have varied in height from 20 to 80 feet. An earthquake on the island of Hondo, Japan, in 1854 was accompanied by a wave which rose 30 feet above the usual level of the water. In 1896 another disturbance on the same coast generated three

waves, the largest of which was 50 feet high on the shore. Ships were torn from their anchorage, and one two-masted schooner was washed nearly a third of a mile inland. The Messina earthquake of December 28, 1908, produced waves which rose nearly 30 feet high on some of the adjacent coasts. Following the eruption of Krakatoa in 1883 waves of enormous height wrought destruction over great distances. On the southern end of Sumatra one wave was over 70 feet high, and carried a gunboat two miles inland where it was left 30 feet above sealevel. In Katimbong the wave rose 80 feet, and on the shallow shore of Merak, on the Java coast, reached the enormous height of 115 to 135 feet[153].

It is evident that such great waves must leave many records of their presence far above the normal level of the sea. Not only large vessels and smaller boats, which readily attract the popular attention, but fish and other forms of marine life are left stranded far inland and high above the reach of the highest tides or greatest storm waves. The bones of whales, and well-preserved marine shells occasionally found high above the sea, must not too readily be accepted as proof of a very recent uplift of the land. Successive earthquake waves in a given ocean may deluge the coasts of all the surrounding continents; and we must therefore expect to find driftwood, shells, and bones of fish well above sealevel at occasional points on almost any shore.

TIDAL WAVES

The great periodic motion of the sea known as the *tide* combines some of the features of oscillatory waves with others belonging to waves of translation. It has been described by Russell as a "compound wave of the first order" (wave of translation) having more of the characteristics of waves of this order than of oscillatory waves[154]. There is no necessity, however of our entering into a discussion of the origin and character of the tidal wave, since the only elements of its motion of vital interest to the student of shore forms are the currents it produces, and the height to which it rises; both of which points are considered in another part of this volume.

Wheeler has expressed the belief that the rising and falling of the tide is accompanied by the production of "tidal wavelets"

which are not the result of wind action, but are in some way genetically related to the tide itself[155]. The explanation of their origin which he gives is not wholly satisfactory, and his theory seems to be based upon his observation that waves from 6 to 24 inches in height break upon the beach at the rate of ten to twenty a minute " when there is an entire absence of wind or other disturbing cause." In the absence of sufficient evidence to connect such wavelets with the tides, we may perhaps more safely regard them as due to the action of gentle breezes and occasional gusts of wind, possibly some distance away, which even on the calmest day never permit the ocean surface to become absolutely quiescent. Haupt[156] states that the flood tide produces waves which break obliquely on the beach, and speaks of " the angle at which the flood breaks upon the shore." But since he also speaks of these supposed tidal waves as " breakers racing along the shore," and quotes Mitchell's description of the manner in which the " larger class of swell or rollers " strike the shore as an example of tidal wave activity, it would appear that Haupt has mistaken the ground-swell of distant storms for tidal waves. A similar misapprehension may have been responsible for Marsh's curious idea that " on most coasts the supply of sand for the formation of dunes is derived from tidal waves," since " the momentum acquired by the heavy particles in rolling in with the water tends to carry them even beyond the flow of the waves "[157].

STANDING WAVES; SEICHES

Under certain conditions there may exist oscillations of the water known as standing waves, in which the water particles do not describe closed orbits, but return through the same paths by which they advance (Fig. 8). The surface water moves upward in all of the crest, and downward in all of the trough, and the vertical movement of the particles is at a maximum under the crest where the horizontal movement is nil[158]. Horizontal movement is at a maximum under the nodal lines (Fig. 8). An example of the standing wave is the *seiche*, typically developed in inland lakes, and extensively studied in Lake Geneva by Forel. This movement consists of a periodic rise and fall of the water surface which is initiated by winds piling up the water at one end of the lake, by sudden variations in atmospheric

pressure, by earthquakes, by landslides, or by some other disturbance; and which continues for some time with gradually diminishing intensity. Each body of water has its own period, appropriate to its dimensions, and the extent to which the water rises and falls depends on the dimensions of the water body and the nature of the disturbing force. The principal seiche on Lake Geneva has an amplitude of from 8 centimeters to 2 meters[159].

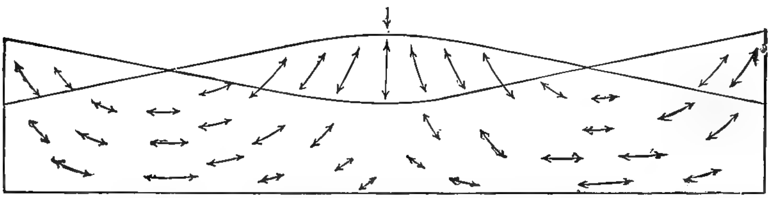

FIG. 8. — Diagram to illustrate the movement of water particles in standing waves, such as the seiche.

Seiches also occur along the coasts of the ocean, especially in bays and straits. Examples of these and other types of seiches are described in Harris's "Manual of Tides"[160], and Thoulet's "Océanographie Dynamique"[161]. According to Dawson[162] a seiche at Yarmouth, Nova Scotia, had an amplitude of from 128 to 143 centimeters, or a maximum change of level of nearly 5 feet. As a rule, however, most seiches have an amplitude of a few inches only. The period varies from a few minutes in small water bodies to many hours in large ones, and the velocity of the water particles participating in the oscillation is not great. Indeed, the direct effect of seiches upon shoreline processes is probably almost negligible. In the rare cases where the amplitude is great the effect of seiches is temporarily to raise the zone of ordinary wave activity to an appreciable extent; and occasionally the rising and falling of the water will cause currents of some importance through narrow straits or inlets; but these are exceptional cases and do not justify us in devoting further space to the subject of seiches in this connection. A good account of this type of wave motion, with a short bibliography, will be found in the work by Harris already referred to, while Darwin's volume on "Tides and Kindred Phenomena" gives a description of Forel's important researches and a list of his classic papers[163].

BOUNDARY WAVES

Where a layer of lighter surface water overlies a heavier water stratum, any sudden wind which creates or accelerates movement of the surface water will cause a rise of the underlying heavier water at the point affected, and a corresponding depression in the heavier water farther forward. The development of such "boundary waves" at the plane of contact of two liquids having different densities can readily be demonstrated by repeating Sandström's experiment, in which one of the layers was colored in order to distinguish it from the other, and a local air current was artificially generated[164]. (Fig. 9.)

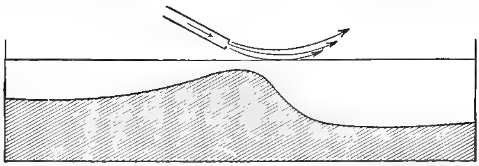

FIG. 9. — Boundary wave formed by local air current over liquids of different densities. (After Sandström.)

When fresh water from some large river flows out over the heavier salt water of the sea, conditions favoring the formation of boundary waves exist. Such waves move very slowly, their velocities depending upon the difference in density of the two water layers and increasing with the square root of this difference[165]. If the generating wind cease, the boundary waves advance to the margins of the containing basin, where they are partially destroyed and partly reflected back beneath the surface. In their progress they give rise to surface waves of the same length, but much smaller height, the crests of the surface waves being directly above the troughs of the boundary waves[166]. Since boundary waves have a low velocity, and the water particles involved move still more slowly, it may be doubted whether they

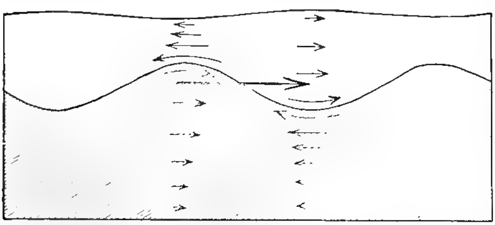

FIG. 10. — Diagram showing movement of water particles in overlying fresh water (white) and underlying salt water (shaded) during the passage of boundary waves from left to right (long arrow). (After V. W. Ekman.)

are of importance in shore processes. A good brief summary of the character of these waves is published by Helland-Hansen and Nansen in their report on the Norwegian Sea[167], while the mathematical theory applicable to them has been developed by Stokes[168].

RÉSUMÉ

In the foregoing pages we have gained some idea of the nature of that force which is the most important agent in the modeling of shore forms. We have considered the form and characteristics of waves on the deep sea in order that we might the better appreciate the changes which they undergo as they approach the coast and begin their geological work. The motion of the water particles in different types of waves; the nature of wave motion in deep water and over shallow sea-bottoms; the origin of storm waves, swells, and surf; the magnitude of waves and the conditions which govern their size; and the velocity of wave advance in both deep and shallow water, have in turn received our attention. With these points in mind we are prepared to enquire into the energy expended by waves upon the shore, and the work thereby accomplished.

In spite of the apparently hopeless chaos presented by the surface of a stormy sea, we know that the waves are controlled by definite natural laws, and that the different elements of form and motion are in systematic relation to one another. So perfect is this relationship that one may stand upon the beach and time the breakers as they dash themselves to pieces at his feet, and learn thereby the length and velocity which these same waves had, hours ago, far away upon the deep sea. On the other hand, we know that not all the laws which control the behavior of waves have been discovered; and we have seen that where different types of waves act simultaneously upon the same water body it may be difficult or even impossible to analyze the resultant movements of the water. We are therefore prepared to find that through the work of waves upon a coast the shoreline is changed according to definite natural laws which are in part, at least, discoverable. But we shall not be surprised if, in the present state of our knowledge of waves, we find it impossible to explain all of the changes which take place upon a shore under their influence.

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CHAPTER II

THE WORK OF WAVES

Advance Summary. — Water waves, whose general characteristics were discussed in Chapter I, possess energy capable of effecting profound changes upon the margins of the land or upon artificial structures with which they may come into contact. Geologist, geographer, and engineer must each concern himself with the nature and magnitude of wave energy, and with the manner in which waves accomplish their work. The layman finds the destructive energy of waves a source of interest and wonder, and not unnaturally regards the meeting-place of land and sea as one of the most fascinating of Nature's laboratories.

In the present chapter the nature of wave energy is first discussed, and the manner of wave attack upon cliffs and sloping shores is briefly treated. It is then shown that the dynamic pressures exerted by waves may be measured with reasonable exactness, and calculated and measured pressures are shown to be in substantial agreement. Some of the most striking examples of damage done by storm waves are next passed in review, in order that the reader may visualize the magnitude of the force responsible for the modification of shore features and the manifold methods of its working. In order to determine which parts of a shore or what artificial structures will suffer most from wave attack, it is essential to know precisely what factors control wave energy, and these are briefly considered. A process of "wave refraction" is shown to be responsible for the concentration of wave attack upon projecting headlands and for the comparative immunity of shores about the heads of bays. In conclusion, attention is directed to the vitally important question as to how far below the water surface wave action may be appreciable.

Wave Energy. — It can readily be shown that a wave transmits energy along the surface of a water body, and delivers this energy on the beach or against some artificial obstacle. When a ship is

propelled through the water, a wave is pushed up by the bow. It took a certain amount of energy to raise this mound of water, and that amount was taken away from the energy of the moving vessel, thereby causing the vessel's motion to be retarded. The wave passes over the surface of the water; and if it finally dashes upon some beach, the energy there expended is the same energy imparted by the moving boat, less a small amount lost through friction.

The mere spreading apart of the water by a vessel's bow does not require the expenditure of energy. If it were not for other causes of resistance, a ship once started through the water would move on forever, unimpeded by the pushing apart of the water in front. The common idea that a vessel's bow is made sharp so that it may cut into the water like a wedge and more easily push it out of the way, is erroneous. No part of the resistance to a ship's motion arises directly from the pushing of water to either side by the bow[1]. A great deal of resistance does arise, however, from the fact that energy is used up in making waves, and one object of the naval architect is to design a vessel of such form that it will produce the fewest and smallest waves possible.

The energy of a wave depends upon its length and height, and is of two types: the kinetic energy due to the orbital movement of the water particles; and the potential energy due to the fact that the center of gravity of the mass of water composing a wave is raised slightly above the position it occupies when the water is at rest. It can be shown that the two types of energy are exactly equal in amount; in other words, the energy of a wave is half kinetic and half potential. Since we know that a cubic foot of sea water weighs about 64 pounds, it is easy to calculate the total energy of either shallow-water or deep-water waves in foot-tons per linear foot of wave crest. The formulæ employed in such calculations are too complex for discussion here, but may be found in Gaillard's treatise[2], and similar works.

During the advance of a deep-water oscillatory wave one-half of the total wave energy is transmitted forward with the wave form. The energy of shallow-water oscillatory waves is from 1 per cent to 11 per cent less than the energy of deep-water waves of equal length and height, but just as in the case of deep-water waves one-half the total wave energy is transmitted onward. In both cases it is the potential energy which is thus carried

forward with the wave[3]. In the wave of translation the energy is also partly kinetic and partly potential; but as this wave leaves still water behind it, at the original level, the entire wave energy must pass forward with the wave form. We have already seen that when oscillatory waves pass into water which shoals very gradually, they are slowly transformed into waves of translation, or at least acquire some of the characteristics of such waves. From this it follows that an oscillatory wave may, by changing into a wave of translation, deliver at the shore all, or nearly all, its energy[4]. This may help to explain the fact that the blows of storm waves against a cliff or sea wall often exceed in violence the available energy calculated for the waves on the assumption that they are waves of oscillation.

Nature of Wave Attack. — The nature of the force exerted by a wave upon any obstacle, such as a cliff or beach, depends in part upon the type of wave and its condition at the moment of collision with the obstacle. If an unbroken oscillatory wave strikes a vertical wall or cliff the base of which reaches down to deep water, the wave is reflected back. At the instant of contact the crest of the wave rises to twice its normal height and the cliff is subjected to the hydrostatic pressure of this unusually high water column. The absence of any forward thrust of the water mass under these conditions is shown by the behavior of boats which have been observed to rise and fall with successive waves without touching the vertical wall only a few feet distant. Hagen[5] concludes that under such circumstances débris must accumulate at the base of the wall and that therefore the prejudice against vertical sea walls and harbor walls, based on the fear of undermining by wave action, is ill-founded.

A wave of translation striking a vertical wall or cliff under the same circumstances is also reflected; but it delivers against the cliff a vigorous push due to the forward thrust of the whole mass of the wave, in addition to subjecting the obstruction to hydrostatic pressure. Stevenson[6] found that " oscillatory waves become waves of translation when they reach the unfinished part of a vertical sea wall, and that they then exert a force nearly 6 times greater than if they had remained waves of oscillation." If either type of wave breaks just before reaching the cliff, in such manner that the forward falling crest of the wave strikes the cliff face, the only force exerted is that due to the forward motion of the water

particles. This motion may exceed the velocity of the wave itself at the time of breaking, the crest shooting forward beyond the main body of the wave as it falls. When an oscillatory wave breaks a short distance out in front of the cliff, so that the forward pitching crest does not strike the cliff, but plunges into the water at its base, the regular orbital motion is destroyed and a " whirlpool turbulence " is produced, the forces of which are not easily analyzed. In a similar manner, if a wave of translation breaks just before reaching a cliff, it " becomes a surge or broken foam, a disintegrated heap of water particles, having lost all continuity." The moving waters of the surge or whirlpool turbulence may exert considerable dynamic pressure on the base of the cliff, and some hydrostatic pressure, depending on the height to which the water rises. When either oscillatory waves or waves of translation break far out from the base of the cliff, smaller waves of translation may traverse the intervening water and operate upon the cliff in the manner already described.

On a sloping shore of fairly steep inclination, oscillatory waves may arrive almost at the beach before losing their essential characters. When such a wave breaks the falling crest dashes down upon the water which is returning seaward from the swash of the preceding wave. The falling wave crest thus strikes a cushion of moving water which may be of considerable thickness. A zone of great confusion is thus produced, the force of the wave is largely dissipated, and part of its volume augments the sheet of water moving seaward, while a larger part starts up the beach. Almost instantly the remainder of the breaking wave overtakes the zone of disturbance, the forward oscillation under the crest checking and possibly reversing the seaward motion of the bottom water, while the landward moving water is enormously augmented in volume. At the same time the orbital motion of the water is largely destroyed, and in the form of a confused mass it rushes up the beach until stopped by gravity and friction, when it flows back with gradually increasing velocity to meet the next oncoming wave. Under these conditions much of the energy of the wave is consumed by friction in the turbulent waters, while another part is expended in the impact of the falling crest upon the bottom wherever the sheet of seaward moving water is effectively pierced. The beach itself is affected

PLATE IV.

Photo by A. M. Cromack.

Storm wave striking face of sea wall at Scarborough, England.

mainly by the sheet of water which is propelled up the slope by the remnant of the wave's energy of motion, and which returns under the action of gravity.

It should be noted that after the oscillatory wave breaks, the confused mass of water propelled up the slope of the beach may be regarded as an irregular type of wave of translation. When a typical wave of translation breaks immediately at the foot of the beach, its falling crest must also meet the backward flowing water cast up by the preceding wave, and give rise to much the same phenomena as the breaking oscillatory wave.

On a coast bordered by water so shallow that large oscillatory waves are broken some distance out from the shoreline, waves of translation and small oscillatory waves alone may reach the beach. If the beach slopes very gradually under water, there may be a secondary line of surf a short distance out where these waves break, and the amount of wave energy which finally reaches the beach itself may be quite insignificant. On the other hand, if the water between the shoreline and the zone where the great oscillatory waves break is of fairly constant depth, and the shore rises fairly abruptly at the inner margin of the shallow, waves of translation of considerable size may deliver their whole energy upon the beach. The latter is then subjected to the static pressure due to the wave height, and the dynamic force of the rapidly moving water particles.

When a wave comes in contact with a vertical or very steep wall or cliff, a relatively small portion of the wave mass may be shot upward (Plates IV and V). It appears that under these circumstances the energy of a large portion of the wave is suddenly communicated to the smaller water mass. The result is that the velocity of this mass may be very great, and it may deliver a blow of terrific force upon a small area. Overhanging cliffs or projections from cliff faces, the roofs of sea caves, and other masses of rocks favorably situated may be subjected to blows from below which have the strength of a battering ram. The energy expended is the kinetic energy due to the swift motion of the water masses.

Masses of water shot into the air in the manner just described may encounter no obstacle in their upward flight, but may descend upon the level summit or sloping face of a cliff, the surface of a beach, or some artificial structure. Such falling masses

PLATE V

Photo by A. M. Cromack.

Water forced vertically upward by wave breaking against sea wall at Scarborough, England.

of water are capable of executing considerable damage because of the great energy they acquire by descending with the ever-increasing velocity due to gravitation.

Wave Dynamometer. Stevenson has shown that the action of a wave is not at all like the sudden impact of a hard body, but is analogous to the steady pressure of a current, because the wave acts with a continuous pressure for an appreciable length of time[7]. It follows from this that if waves are allowed to come against a vertical plate which has a spring back of it, and if the

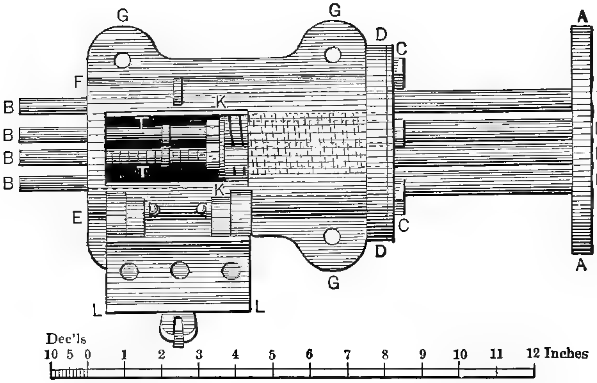


Fig. 11. — Stevenson's Wave Dynamometer.

DEFD is a cast-iron cylinder, bolted to the rock by the flanges at *G*. *AA* is an iron disk against which the waves impinge, fastened to guide rods *BB*, which pass through holes in the plate *CC*. When waves strike the disk *AA*, rings of leather *TT* are moved along the guide rods, registering the extent to which the spring is lengthened. *LL* is a door opened for the purpose of reading the instrument.

change in length of the spring due to the pressure against the plate is determined, we shall have a proper measure of the dynamic pressure exerted by the wave. Stevenson[8] devised such an instrument, called a dynamometer, and was the first man to measure the force of waves. Gaillard confirmed Stevenson's results, but pointed out that the spring dynamometer measures only the dynamic pressure of the moving water in the wave, and gives no information as to the static pressure resulting from the weight of the water mass. This is due to the fact that static pressure is just as great on the back of the plate as on the front, and therefore produces no effect on the spring. He therefore designed a dia-

phragm dynamometer having a sheet of rubber stretched over one end of a short iron cylinder, the other end being closed by an iron plate. Pressures due to the waves thus affect but one side of the instrument by pushing in the rubber diaphragm, and the magnitudes of the pressures are determined by means of a gauge attached to the cylinder. To measure the static pressure due to the column of water in the wave the instrument is placed with its face horizontal and upward at the desired depth in the water. When placed with its face vertical so as to receive the full impact of the advancing wave the instrument records both dynamic and static pressures[9].

Measurements of wave force with dynamometers indicate that the static pressures exerted by waves are considerably less than their dynamic pressures. On Lake Superior, Gaillard found the static pressure of a wave 10.5 feet high and 150 feet long, to be 3.23 lbs. per square inch, or about 450 lbs. per square foot, the dynamometer being 9 feet below the wave crest. The dynamic pressures of waves 10 feet high and 150 feet long varied from 460 to 965 lbs. per square foot on a dynamometer placed about a foot higher than that for the measurement of static pressures[10]. Adequate observations of wave pressures by means of suitable dynamometers have not yet been made, those for static pressures being especially insufficient in number.

Measurements of Wave Energy. — In order to gain some conception of the enormous power of waves we have only to consider the theoretical pressures calculated for waves of different size, the actual pressures recorded by dynamometers on exposed coasts, or the damage to harbor works done by storm waves. Gaillard has calculated that a wave 10 feet high and 100 feet in length may strike an obstruction with a pressure of 1675 lbs. per square foot, while a wave 12 feet high and 200 feet long should exert a maximum dynamometer pressure of 2436 lbs. per square foot. The total theoretical energy of such a wave is 109 foot-tons for every linear foot of wave crest. Great ocean waves such as those which destroyed part of the breakwater at Wick, Scotland, in 1872, if we assume a height of 42 feet and a length of 500 feet, should produce a pressure of 6340 pounds per square foot[11].

Dynamometer readings show that during storms on Lake Superior the waves develop a force of from 1600 to 2500 lbs. per square foot[12]. Stevenson found that the Atlantic Ocean

PLATE VI.

Marine cliff at Highland Light, Cape Cod, rapidly retreating under wave attack. It has been estimated that a strip of coast two miles broad may already have been cut away by the waves, and that in eight or ten thousand years all the land between the Atlantic Ocean and Cape Cod Bay may be consumed.

waves near the island of Tyrec on the Scottish coast had an average force of 611 lbs. per square foot during the summer months, whereas the average for the winter months was 2086 lbs., or more than three times as great. The greatest force recorded at this point was 6083 lbs. or practically that calculated theoretically for a very large ocean wave. Another reading of 5323 lbs. was secured. On the east coast of Scotland pressures of more than 6000 lbs. per square foot were recorded[13].

Damage by Storm Waves. — Such enormous pressures are capable of producing remarkable results. Stevenson describes an instance in which a block of stone weighing $7\frac{1}{2}$ tons and situated 20 feet above sealevel was driven before the waves for a distance of 73 feet over rugged ledges[14]. At North Beach, Florida, a solid block of concrete weighing 4500 lbs. was moved 12 feet horizontally and turned over on its side, while a second block weighing 3600 lbs. and having its center at high-water level was shifted several inches by waves which were not over 4 feet high. During a severe storm on December 25, 1836, stones forming part of the breakwater at Cherbourg and weighing nearly 7000 lbs. were thrown over a wall 20 feet high which surmounts the stone embankment. In the harbor of Cette a block of concrete 2500 cubic feet in volume and weighing about 125 tons was shifted more than 3 feet from its original position. Perhaps the most wonderful example of wave work is that accomplished by ocean storm waves upon the breakwater at Wick in December, 1872, and described in Stevenson's treatise on "Harbors." The seaward end of this breakwater was protected by a monolithic block of cement rubble 45 feet long, 26 feet wide and 11 feet thick, and weighing more than 800 tons, resting on great blocks of stone which were bound solidly to the monolith by iron rods $3\frac{1}{2}$ inches in diameter running through holes in the stones and embedded in the cement rubble. The entire mass, weighing 1350 tons, was torn from its place by the waves and dropped inside the pier, where it was found unbroken after the storm subsided. A much larger mass of concrete was substituted for the one removed, the new block having a volume of 1500 cubic yards, and weighing 2600 tons. In 1877 this enormous mass was similarly carried away by the waves[15].

The terrific impact which a wave may deliver against the face of a vertical wall may be appreciated from the fact that the

facing stones of the Wick breakwater, having the same density as granite, were shattered by the sea in February, 1872. At Dunkirk waves from the narrow southern arm of the North Sea strike the coast with an impact which causes a trembling of the ground more than a mile inland[16]. That waves have the power to wrench from place objects situated well above the main body of the wave is shown by the effects of a storm upon Dhuheartach lighthouse on the west coast of Scotland, during which fourteen stones weighing 2 tons each were torn from their positions 37 feet above high tide, and dropped into deep water. Cast-iron lamp posts on the pier heads at Duluth, located 19 feet above lake level, have repeatedly been broken off by wave action.

The lifting power of waves is often illustrated by damage to harbor works. At North Beach, Florida, a block of concrete weighing $10\frac{1}{2}$ tons was lifted vertically upward three inches by the wave pressure transmitted through crevices below the mass. During a storm on Lake Superior a mass of trap rock 2 cubic yards in volume and weighing about $4\frac{1}{2}$ tons was raised by a wave from its place alongside an old breakwater at Duluth and deposited on the surface of the breakwater some 5 or 6 feet above its original position. A more striking example occurred at Ymuiden on the coast of Holland, when a concrete block weighing 20 tons was lifted 12 feet vertically by a wave and deposited on a pier above high-water level.

Waves deflected upward by a sloping surface may accomplish work at high levels. The keeper of Trinidad Head light station, on the Pacific Coast, reports that during the storm of December 28, 1913, the waves repeatedly washed over Pilot Rock, 103 feet high. One unusually large wave struck the cliffs below the light and rose as a solid sea apparently to the same level at which he was standing in the lantern, 196 feet above mean high water, the spray rising 25 feet or more higher. The shock of the impact against the cliffs and tower was terrific, and stopped the revolving of the light. Lake Superior waves reached the door of a light-keeper's dwelling situated 140 feet back from the water and $60\frac{1}{2}$ feet above it, carrying away a board walk and doing other minor damage. On the Bound Skerry in the Shetland Islands blocks of stone from 6 to 13 tons in weight have been forced from their places at a level which is 70 to 75 feet above the sea.

The destructive power of the masses of water hurled to remarkable heights by breaking waves is greater than one might suppose. At the Bell Rock lighthouse in the North Sea a ground-swell, without the aid of wind, drove water to the summit of the tower 106 feet above high tide, and broke off a ladder at an elevation of 86 feet. A bell weighing 3 cwt. was broken from its place in the Bishop Rock lighthouse, 100 feet above high water mark, during a gale in 1860; and at Unst, in the Shetland Islands, a door was broken open at a height of 195 feet above the sea. The keeper of Tillamook Rock lighthouse, on the coast of Oregon, reports that in the winter of 1902 the water of waves was thrown more than 200 feet above the level of the sea, descending upon the roof of his house in apparently solid masses. In October 1912, and again in November 1913, the panes of plate glass in the lantern of this same light, 132 feet above mean high water, were broken in by storm waves.

Great damage may be accomplished by the falling water. According to Shield[17] "it is no uncommon occurrence for storm waves, striking a vertical breakwater face, to throw heavy masses of water to a height of at least 100 feet, often very much higher. Such water in its descent on reaching the roadway of the breakwater upon which it falls, will have attained a velocity of about 80 feet per second, or nearly double the velocity and four times the force of the water striking the face of the breakwater." During a severe gale at Buffalo in December, 1899, seventy big timbers, 12 × 12 inches in thickness, 12 feet long, and 10 feet between supports, were broken in two in the middle by the impact of the falling water. This same breakwater was further damaged a year later when waves breaking against it were hurled from 75 to 125 feet into the air, the falling water crushing the big timbers on which it fell as though they had been pipestems.

A part of the geological work accomplished by waves is due to the direct pressure exerted upon air and water imprisoned in crevices, and another part to the sudden expansion of air in crevices and pore spaces when the rapid retreat of a wave creates a partial vacuum outside. The effect of compressed air may be inferred from the fact that waves coming against a breakwater in Buffalo harbor produced such high pressure upon the air under the concrete shell that four circular plates of concrete, 3 feet in diameter, 6 inches thick and weighing 530 lbs. each, serving as covers to

manholes, were lifted from their places. A block weighing 7 tons in the face of the breakwater at Ymuiden was started forward out of its place during a gale, the movement being *toward* the waves which were coming against it. According to Gaillard this phenomenon was caused " by the stroke of a wave compressing the air in the rear of it"[18], but similar results are produced by expansion due to the formation of a partial vacuum in front. In 1840 a securely fastened door in the Eddystone lighthouse was burst *outward* during the attack of storm waves, the circumstances leading Geikie to conclude that " by the sudden sinking of a mass of water hurled against the building, a partial vacuum was formed, and the air inside forced out the door in its efforts to restore the equilibrium"[19].

Another important factor in the work of waves is the effect produced by stones, logs, blocks of ice, and other objects moving with the waves. It has been well said by Playfair[20] that waves thus armed become a sort of " powerful artillery " with which the ocean assails the land. A large block of ice or a log may concentrate its whole momentum upon a very small area with appropriately great results. Thus, Gaillard has suggested that an exceptionally high dynamometer reading on Lake Michigan (when the instrument showed a pressure twice as great as that recorded in the same locality for a more severe storm and greater than any record for the larger waves of Lake Superior) may possibly have been caused by ice or timber. Large stones may be hurled out of the water with high velocities. At Tillamook Rock on the Oregon Coast, fragments of stones are torn from the cliffs during every severe storm and thrown on the roof of the light-keeper's house, about 100 feet above sealevel. " In December, 1894, one fragment weighing 135 lbs. was thrown clear above this building, and in falling broke a hole 20 feet square through the roof, practically wrecking the interior of the building. Thirteen panes of glass in the lantern were broken during the same storm"[21]. As already noted, this lantern is 132 feet above mean high water. The fog-signal siren horns, about 95 feet above the sea, were partially filled with rocks during the storm of October 18, 1912. The windows of the Dunnet Head lighthouse on the north coast of Scotland, which are over 300 feet above high-water mark, are sometimes broken by stones swept up the cliffs by waves[22].

It is perfectly evident that waves which are armed with cobblestones and enormous boulders must accomplish great erosive work when they beat against a cliff or artificial wall. On the other hand, one must not make the mistake of assuming that waves which are not thus armed can accomplish but little work. It is true that large storm waves may beat against a cliff without removing the barnacles which are attached to its face, and that along the shores of saline lakes calcareous tufa may form on cliffs exposed to the impact of large waves[23]. But this merely indicates that the pressure of the liquid mass is so evenly distributed upon all sides of the strong shell, or of the mineral deposit, that the excess of pressure on any one side is not sufficiently great nor applied with sufficient suddenness to cause rupture. The same waves will wrench great blocks of rock from their places in the cliff face, and drive air and water into joint crevices with such force as to loosen large fragments of the cliff and thus contribute to the disintegration of the whole mass. A force which exerts a pressure of thousands of pounds to the square foot will discover lines of weakness in any natural cliff. Even though all sand, boulders, and other rock fragments were speedily carried out of the zone of wave action, and waves of pure water alone attacked the coasts, shorelines would retreat under wave erosion just as surely as they do when the waves are armed with abrasive materials, although the process would certainly go on much more slowly.

British geologists have long appreciated the tremendous power of the waves in destroying land areas, and with good cause; for no part of the British Isles is far removed from the sea, the wave attack on much of the coast is remarkably vigorous, and abundant ancient records and surveys permit careful computation of the rate of cliff retreat at many points. Old maps of Yorkshire show the location of many towns and villages which have been swept out of existence by the waves, their former sites being now represented by sandbanks far out in the sea. In 1829 there was in the harbor of Sheringham, according to Lyell[24], a depth of 20 feet of water where only forty-eight years before had stood a cliff fifty feet high with houses upon it. For over half a century the cliff at Happisburgh retreated at the rate of 7 feet per year, while the cliff between Cromer and Mundesley was cut back 330 feet in the twenty-three years previous to 1861 making an

PLATE VII.

Shakespeare's Cliff near Dover, England. A marine cliff being actively cut back in chalk hills by vigorous wave action.

annual retreat of 14 feet. Matthews[25] estimates that the rate of cliff erosion on the Holderness coast of Yorkshire varies from 7 feet per year in some places to 15 feet in others, while the retreat between Cromer and Mundesley since 1861 is said to have been 19 feet annually. At Southwold the annual rate has varied from 15 to 45 feet. Shakespeare's cliff (Plate VII) near Dover is so vigorously undermined that great landslides descend from the upper part of the cliff, the débris projecting far into the sea until the waves remove it and renew their attack on the cliff base. Such a landslide in 1810 caused a marked earthquake at Dover. Detailed accounts of the rates of cliff erosion about the British Isles will be found in Lyell's "Principles of Geology"[26], while Matthew's "Coast Erosion and Protection"[27] gives more recent data on this question. Further details are abundantly set forth in the reports of the Royal Commission on Coast Erosion of Great Britain[28].

The large blocks of rock dislodged from cliff faces, as well as smaller fragments, are churned together by the waves so long as they remain within reach, either upon the beach slope or in shallow water. The surf zone has been likened by Shaler[29] to a great mill in which angular fragments are quickly rounded and everything in course of time is reduced to the size of sand or fine silt and swept out to sea. How effective is this mill may be inferred from the fact that angular fragments of granite from quarries on Cape Ann, Massachusetts, become fairly well rounded by wave action in a single year, while under favorable circumstances "the wear upon the pebbles amounts on the average to several inches per annum"[30]. On a stormy day the roar of grinding masses of boulders often rises above the sullen thundering of the surf.

A vivid picture of the working of the "sea mill," which grinds great boulders to sand and fine mud, is given by Henwood[31] in an account of the visit made by him to a mine in southwest England which extended out under the sea: "When standing beneath the base of the cliff, and in that part of the mine where but nine feet of rock stood between us and the ocean, the heavy roll of the larger boulders, the ceaseless grinding of the pebbles, the fierce thundering of the billows, with the crackling and boiling as they rebounded, placed a tempest in its most appalling form too vividly before me to be ever forgotten. More than once

doubting the protection of our rocky shield we retreated in affright; and it was only after repeated trials that we had confidence to pursue our investigations."

Conditions Affecting Wave Energy. — We have already seen that the dimensions of waves vary with differences in depth of water, strength and duration of wind, and length of fetch of open water. It follows from this that the amount of wave energy delivered against a shore will vary with these same factors. A coast bordered by off-shore shallows escapes the most powerful wave attack, because large waves cannot traverse the shallow water. Other things being equal, that part of a shoreline facing the greatest stretch of open water will receive the largest amount of wave energy. But it must be remembered that the prevailing winds may come across a shorter stretch of open water, with the result that what appear to be the less exposed parts of a shore may really suffer the more vigorous attack. Account must also be taken of the fact that the *prevailing* wind may not be the *dominant* wind; for a few great storms from one direction may more than offset the effect of long-continued wave attack from the direction of the prevailing wind. It is, therefore, not always a simple matter to determine which parts of a shore will suffer most from the energy of waves. The observer must carefully consider the inclination of the off-shore slope; the depth of water both near the shore and farther out; the presence or absence of shallows; the directions of the greatest stretch of open water, of the prevailing winds and of the greatest storm winds; and a number of other factors which may enter into the case; and must skilfully weigh the relative importance of each factor in a given case before he can reach a safe conclusion.

Among the factors affecting the energy with which waves attack a shore are two not previously mentioned. These are tidal currents and the angle at which the waves meet the shoreline. When a wave encounters an opposing tidal current, the velocity and length of the wave are decreased, the height is increased, and the wave may break much as it would on a shelving beach. A swiftly moving tidal current may thus be quite as effective as a shallow in causing large waves to break before reaching the shore. The south coast of Shetland is protected from the waves of a southwest storm so long as a rapid tidal current off the coast is running, no matter how rough may be the

PLATE VIII.

Wire fence undermined by wave attack and left hanging in mid-air. The road in front of the house has been completely cut away. Third Cliff, Scituate, Massachusetts.

outside sea; but as soon as the current ceases the surf breaks on the shore with great force[32]. On the other hand, if the current is located immediately at the shore, and especially if it flows in the direction of wave advance, the destructive power of the waves may be augmented. Stevenson is of the opinion that the violence of the surf at Whalsey and Wick in northern Scotland is in part due to the action of strong tidal currents; and at certain other places the surf seems to be most destructive when the tidal currents are strongest[33].

The angle at which the waves meet the shoreline has an important effect upon the energy of wave attack. Waves are most destructive when they come in at right angles to the shoreline, and a very slight amount of obliquity materially decreases their power. Those portions of the breakwater at Wick which are assailed by waves coming "dead-on" have suffered much greater damage than other portions where the waves arrive at a slightly oblique angle[34]. It is therefore evident that where the direction of greatest fetch of open water makes an oblique angle with the shoreline, waves from that direction may be less destructive than waves developed on a shorter stretch of open water but approaching the land at right angles to the shore. It must not be supposed, however, that waves approaching a coast from a given direction maintain that direction until they break upon the beach. On the contrary, there is a very marked tendency for every wave to change its direction in such manner as to make its crest parallel with, and its direction of advance at right angles to the shoreline. Inasmuch as this tendency has an important effect upon the development of shorelines, we must give it some further consideration.

Wave Refraction. — When a wave (*ad*, Fig. 12) advances toward a coast, the direction of advance is always at right angles to the wave crest. Nearing the coast, the wave encounters shallower water off the headlands than opposite the bays; and since the velocity of shallow-water waves decreases with decreasing depth, those parts of a wave opposite headlands will lag behind the parts opposite bays, and the wave crest will begin to curve (a^1d^1) in conformity with the curves of the shoreline. If the headlands and bays are not too pronounced, and if the shallowing of the water is not too abrupt, by the time the wave has reached the position a^3d^3 it will have so adjusted itself as to bring its crest

at all points nearly parallel to the shoreline. Or, as Harrison[35] has expressed it, "the velocity of the part which first reaches the shallow being lessened, the whole wave wheels round, and breaks nearly at right angles on the beach." This process of "wave refraction," as Davis has called it, accounts for the fact that swells from distant storms ordinarily are nearly parallel to

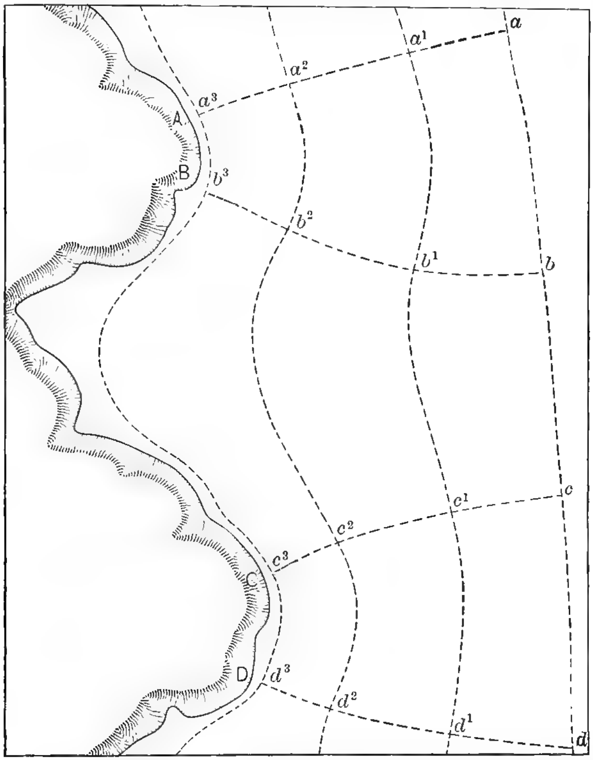


FIG. 12. — Diagram to illustrate the process of wave refraction, whereby wave attack is concentrated on headlands. (After Davis.)

the shore when they break, no matter what may have been the direction of the storm. Even within the narrow limits of a single curved beach an observer may note the tendency of the surf to break directly on shore throughout its length, although the beach may describe an arc of 90 degrees or more.

An important consequence of wave refraction is the concentration of wave energy upon headlands. Since the direction

of wave advance is always at right angles to the crestline, and the latter becomes curved to conform with the curvature of the shoreline, it follows that a large proportion of the waves will be refracted toward the headlands. In Fig. 12 it is apparent that all that portion of the wave between *a* and *b* will be concentrated upon the short stretch of shoreline, AB, on the headland; whereas the part of the wave between *b* and *c* will be distributed over the great stretch of the bay shore, BC. In other words, wave refraction causes an enormous concentration of wave energy upon headlands and a dissipation of energy in bays. The observer who wishes to witness the most sublime manifestations of the power of the sea must seek the exposed headlands of the coast; while the mariner finds comparative safety within the limits of the bays, even where these are broadly open to the sea.

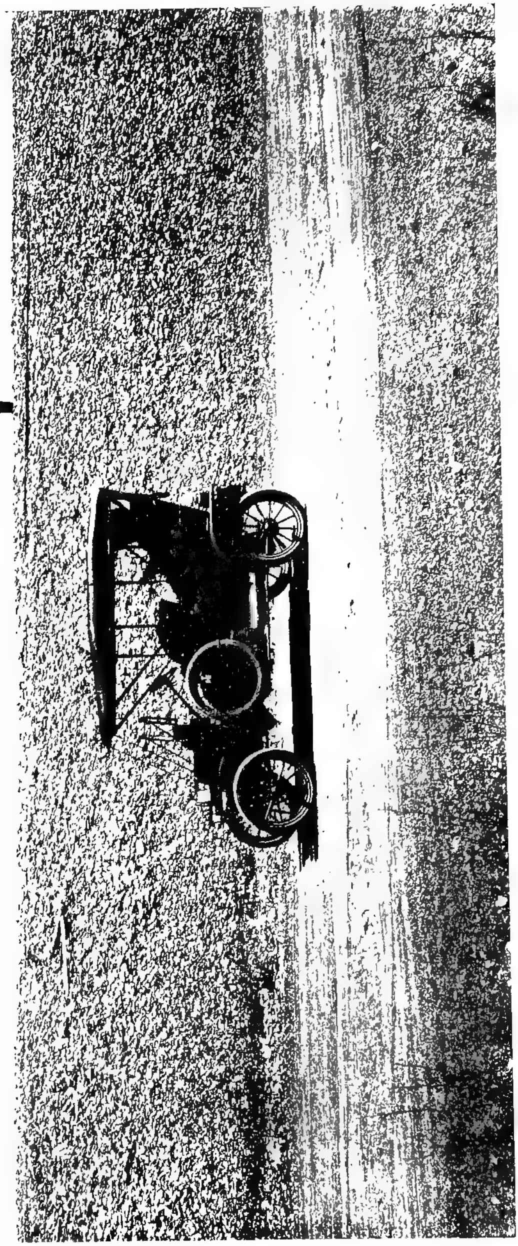
Waves do not always break parallel to the shore. In the first place, no wave can be refracted with sufficient abruptness to render its crest parallel to the sharp and complex irregularities of some shores. In the second place, the water is very deep close to some shores, and wave refraction does not begin to take place until the wave has practically reached the headlands. The wave then breaks against these projecting points of the coast first, and its remaining portions, being imperfectly refracted, sweep upon the shore from the headlands inward at an oblique angle. Furthermore "forced waves," or those which are still being driven forward by the wind which formed them, are not so readily refracted as "free waves" which have passed beyond the limits of the storm. It is for this reason that storm waves are more apt to strike the shore at an oblique angle than are the groundswells which arrive during calm weather. The more perfect refraction of the groundswells is due not alone to the absence of the wind's impelling force, but probably also to the fact that they extend to greater depths and hence are the sooner affected by the refracting influence of a shallowing bottom.

Depth of Wave Action. — The depth to which the ocean waters are affected by waves is a matter of much importance to all students of shore processes. We have already seen that at the depth of one wave length below the surface the water particles of oscillatory waves are moving in orbits whose diameters are only $\frac{1}{535}$ as great as the diameters of the orbits at

the surface. Since the period of the lower orbits is identical with that of the larger surface orbits, it follows that the velocity of the water particles decreases in the same proportion as the diameters of the orbits. In other words, the water particles at the depth of one wave length below the surface move with $\frac{1}{535}$ of the velocity of the surface particles. The ability of oscillatory waves to erode the bottom and to transport débris therefore diminishes rapidly with increasing depth, and soon becomes negligible. In the case of waves of translation the motion of the water particles is theoretically the same from the surface to the bottom, except that the velocity near the bottom should be somewhat less than near the surface, owing to the fact that the water particles there pass through shorter, more nearly horizontal paths in the same length of time. With these theoretical points in mind it will be interesting to inquire into the results obtained by different students of this phase of wave activity, and to review their opinions as to the maximum depth of efficient wave action in nature. Unfortunately, few writers distinguish between the effects of oscillatory waves and waves of translation.

Captain E. K. Calver, R. N., has observed waves which changed their color upon passing into water from 40 to 50 feet deep because of their abrasive action upon the bottom[36]. Sir John Coode studied the movement of shingle in the vicinity of the Chesil Bank on the south coast of England, by descending to a depth of 60 or 65 feet below the surface of the sea in diving dress. He found that after a heavy storm the shingle, which was previously covered with barnacles, was quite free from these shells, proving a movement of the coarse material at a depth of nearly 50 feet[37]. According to Hermann Fol, whose "Impressions d'un Scaphandrier" are vividly recorded in the Revue Scientifique for 1890[38], a diver at a depth of 100 feet is tossed back and forth by the vigorous oscillatory movement of the bottom water whenever groundswells are running on the surface. Hunt quotes the testimony of pilots and masters to the effect that after a wave has broken over a vessel, sand is frequently left on the decks even when the water has a depth of 75 or 80 feet[39], and describes a jar brought up in a trawl from a depth of 220 feet into which gravel the size of a hazelnut had been washed by wave agitation. Robert Stevenson states that fish disappear

PLATE IX.



Cobblestones cast into a high ridge well above sealevel by storm waves. Near Rye, New Hampshire.

from the fishing grounds in the North Sea during storms, due to the agitation of the water by wave action to a depth of 200 feet or more[40]. The same authority notes that at the Bell Rock lighthouse, off the east coast of Scotland, large stones, containing upwards of 30 cubic feet and weighing two tons or more, are often thrown upon the rock from "deep water" by the waves[41]. Thomas Stevenson has made a very interesting comparison between the depths at which mud reposes on the floor of different parts of the North Sea, and the vigor of wave action in those places. He finds that there is a direct relation between these two phenomena, the depth of the level at which mud accumulates increasing in much the same proportion as the violence of the waves. From this we may infer that the upper limit of mud accumulation is a measure of the maximum depth of wave disturbance in a given locality. Applying this rule to the North Sea, we find that in protected areas, as the inner parts of the Moray Firth and the Firth of Forth, and along the Holland coast in the narrow southern part of the sea, wave action reaches to a depth of 25, 50, or 100 feet; while in exposed places the disturbance is appreciable to a depth of from 300 to 500 feet or more[42]. According to J. N. Douglas, the fishermen off Land's End bring up stones one pound in weight, which have been washed into their lobster pots at a depth of 180 feet by the action of the ground-swell, while coarse sand is often washed from a depth of 150 feet by storm waves and hurled to the lantern gallery of the Bishop Rock lighthouse, 120 feet above low-water[43]. Kinahan reports the moving of stones weighing several hundred pounds by wave action in water from 90 to 120 feet deep on the coast of Galway[44].

In contrast to the above records of significant wave action at great depths, may be mentioned a few instances of the inefficiency of wave action a short distance below the surface. At the Cherbourg breakwater blocks of rubble stone 23 to 26 feet below low-water are reported by Wheeler as remaining unmoved in the roughest sea. According to the same authority, the rubble mound upon which the Alderney breakwater was later erected remained three years undisturbed by winter storms below the level of 15 feet below low-water[45]. Indeed, Wheeler goes to the extreme of limiting "the disturbance caused by the formation of waves . . . to a distance below the surface about

equal to the height of the wave"[46]. At Port Elizabeth, in South Africa, Mr. Shield found that blocks of rubble stone, weighing from 1 to 1½ cwt. remained unmoved at a depth of 22 feet when the waves were 15 to 20 feet high[47]. Coode reports that in the same locality the movement of sand on the sea-bottom ceases 20 feet below the surface[48]. Delesse states that submarine portions of engineering structures are seldom disturbed below a depth of 16 feet in the Mediterranean, and 26 feet in the Atlantic[49].

Too much importance must not be attached to the negative results just mentioned. In some of the cases we are not in possession of sufficient information regarding the degree of exposure of the localities in question, or of the size of the waves there generated. Engineering structures and masses of rubble stones may be so keyed together, or may have such external forms, as to receive the shock of vigorous waves without harm, while loose materials on the bottom are at the same time materially affected. High waves of short length will not affect the water to as great a depth as lower waves of greater length. The positive evidence of wave disturbance at depths of several hundred feet is sufficient to prove that however ineffective some waves may be, other waves under favorable conditions will produce an effect in deep water. We may take 600 feet as the limiting depth of ordinary wave disturbance, although Cornish sets 900 feet as the limit for the largest recorded waves[50]. Geikie states that ripple marks are sometimes produced (by waves) in fine sand at a depth of 600 feet, and Airy apparently attributes the breaking of groundswells in water of the same depth to interference with the bottom[51]. Agassiz seems to recognize the possibility of wave action off the coast of Florida to a depth of 600 feet[52]. More definite figures are given by Cialdi, who asserts that large waves will erode the bottom to a depth of 40 meters in the English Channel and Adriatic Sea, 50 meters in the Mediterranean Sea, and 200 meters, or about 650 feet, in the open ocean; and that at such depths the waves will put débris in motion and grind it together[53]. Still more convincing are the results of experiments made by Siau[54] near Saint-Gilles on the Isle of Bourbon, off the coast of Madagascar. This ingenious investigator found that by sounding with a weight well coated with tallow he could determine the presence of ripple marks on the sea-bottom not only because the impression of

the ripples was imprinted upon the tallow surface, but also because the heavy particles concentrated in the troughs and the light particles collected on the crests of the ripples adhered to the tallow in parallel bands. In this way Siau was able to prove the existence of wave-formed ripple marks, and hence of wave action, at a depth of 617 feet. In a letter to Nansen[55] Sir John Murray states that great storms off the north coast of Scotland agitate fine mud at a depth of 600 feet. Murray also quotes Vionnois as authority for the statement that in the Bay of St. Jean de Luz the bottom is agitated during storms at a depth of 300 meters, or nearly 1000 feet[56]. Unfortunately, while these authors evidently refer to oscillatory waves in their discussions, they do not definitely exclude the possibility that waves of translation may be responsible for the deep-water movements. Nor can we be certain, in some of the cases cited, that currents may not have produced the ripple marks and other phenomena attributed to wave action.

There is no theoretical reason, however, why we should doubt the possibility of appreciable oscillatory wave action down to a depth of 600 feet. Observations with the naked eye and with the microscope convinced the Weber brothers that during the passage of oscillatory waves there is some slight movement of the water particles to a depth below the surface equal to 350 times the height of the waves[57]. Accordingly a wave 40 feet high should affect water particles 14,000 feet below the surface. At a depth of but 600 feet this movement must be quite pronounced, despite the rapid decrease in amplitude of oscillation from the surface downward, and notwithstanding that the maximum theoretical depth of wave disturbance may not ordinarily be attained in the ocean because of the long time required for the downward transmission of surface oscillations, which latter may cease or change direction before the lowest water strata are set in motion[58]. Assuming a groundswell with a length of 1350 feet and a height of 16 feet, which is well within the possible limits, the water particles at a depth of 600 feet ($\frac{4}{9}$ of the wave length) would move in orbits having a diameter of 1 foot. The period of such a wave is about 16 seconds; hence the water particles at the depth indicated would oscillate with a maximum velocity of 1 foot in 5 seconds, or .06 meter per second. If at the bottom the path of oscillation were reduced to a straight

line 1 foot in length, the velocity for a wave of the same period would be about .04 meter per second. Such an oscillation would disturb clay, fine mud, and probably the very finest sands. Forbes has shown that fresh water moving in a shallow trough with a velocity of .077 meter per second will stir up moist brick clay[59], while Sorby recently found that a current of 6 inches (.15 meters) per second would drift along common sand grains one hundredth of an inch in diameter and that "the very fine Alum-Bay sand $\frac{1}{300}$ inch in diameter" would be moved by a velocity as low as .04 meter[60]. According to de Lapparent a river with a bottom velocity of .15 meter per second will transport coarse mud[61], whereas Lyell says this same velocity will move fine sand, and Hunt puts the lower limit for ordinary fine sand at .10 meter per second. The foregoing figures are based on observations in shallow fresh water. If we consider the conditions of temperature, pressure, salinity and viscosity which would exist at a depth of 600 feet in the sea, we find that sand particles of a given diameter ought to be moved by a slightly lower current velocity than in the cases cited. It seems reasonably certain, therefore, that with an orbital diameter of 1 foot and a period of 16 seconds there would be appreciable disturbance of the finer deposits on the sea floor. Even were the orbital diameters as small as 1 inch and the period from 10 to 20 seconds, Cornish is of the opinion that the motion of the water would still be sufficient to hinder the deposition of the finest kinds of mud[62]. When associated with slow-moving tidal or other currents, a very gentle oscillatory movement of the water due to wave action may produce an important effect.

In the words of Cornish, "We may say with confidence, as a theoretical inference, that the agitation of wind-formed waves affects the bottom of the sea as far as the edge of the continental platform to such an extent as (in co-operation with tidal and other currents) to keep very fine mud moving about until it has an opportunity of subsiding over the edge of the continental shelf"[63]. On the other hand, it is evident that only the finest material will be affected at such depths, and that erosion of the bottom will be almost imperceptibly slight, so long as oscillatory waves alone disturb the water. Waves of translation are not as common in such deep water as nearer shore, but whenever they do occur we should expect, on theoretical grounds, a velocity of

the bottom water comparable to that at the surface, and therefore capable of effecting noteworthy erosion and transportation. A sufficient body of observed facts to establish this theory is not available. We are reasonably sure, however, that on exposed coasts the sea-bottom is not wholly free from some kind of wave agitation down to a depth of 600 feet at least.

RÉSUMÉ

We have inquired into the origin and character of the energy developed by waves, and have gained some idea of the tremendous power which they may exercise under favorable conditions. It has been seen that natural shores, as well as artificial structures, must suffer severely from wave attack. The conditions affecting the vigor of wave action at the shore have briefly been discussed, and the effects of wave refraction considered more fully. An inquiry as to the depth of wave action has resulted in the conclusion that the sea-bottom is affected by waves to the edge of the continental shelf, or approximately to a depth of 600 feet.

But waves are not the only forces of nature which expend their energy upon shores. Currents of various types play an important rôle in modelling shore forms, and must therefore receive our attention before we proceed to a study of the evolution of shorelines under the combined influence of waves and currents.

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CHAPTER III

CURRENT ACTION

Advance Summary. — Shore débris is subject to transportation by many different kinds of currents. It is the purpose of the present chapter to discuss the more important of these currents and to describe the movements of débris which they cause. Following a brief preliminary summary of the types of currents to be treated, there is presented a detailed analysis of wave currents, and of the profoundly important process of "beach drifting" for which they are responsible. Tidal currents are next considered, and while their value as a factor in beach construction has undoubtedly been much exaggerated, it is shown that they perform a significant function in modifying shores, particularly the shores of estuaries. Currents generated by seiches have but a theoretical importance, except in a very few localities, and therefore receive but scant attention here. Currents caused directly by the friction of winds blowing over water surfaces deserve a more extended treatment. It will be seen that such currents are in some cases of a permanent character, in others purely temporary, while a third group varies in direction or character with changes in the seasons. Some of these "wind currents" are far removed from the lands and consequently play no rôle in shore development; but others locally wash the margins of continents or islands and have their share in shoreline work.

A special class of currents, generated by winds but modified by other causes, comprises the great whirls of the major oceanic circulation, and these are treated separately under the name "planetary currents." They seldom come in direct contact with the lands and are therefore of minor importance to the student of shorelines. Currents due to differences in atmospheric pressure, and convection currents, are likewise shown to play but an insignificant rôle in shore processes. Salinity currents, arising from differences in specific gravity between waters having a different salt content, are well developed in certain

straits, where they may locally control the movements of débris. For this reason currents of this type are treated somewhat fully and special consideration is given to examples at the mouths of the Baltic, Mediterranean and Red seas. River currents and their relation to delta growth are briefly described, and a similar treatment is accorded the "reaction currents" which flow into river mouths under certain conditions. Neither type of current deserves a major place in our discussion. Eddy currents, frequently associated with some of the currents mentioned above, deserve and receive a short space in our text. The important hydraulic currents generated as by-products of various other types of currents are not treated separately, but in association with the movements with which they are genetically connected. The chapter closes with a special consideration of the great complexities of current action.

Types of Currents. — If we define a current as a more or less restricted body of water moving in a definite direction, it is evident that various types of currents may exist in the sea. During oscillatory wave motion, masses of water move first forward, then backward; and in waves of translation there is a forward movement, then a halt, followed by another forward movement, and so on. These short but of repeated movements of the water may affect shore deposits in much the same manner as more continuous currents, and we may therefore speak of them as *wave currents*. They are in many respects analogous to those currents which are associated with the great oscillatory movement of the sea water known as the tide, and which are commonly called *tidal currents*. Bays and straits, as well as lakes have periodic oscillations of their waters called seiches. If these oscillations are of considerable amplitude, the rising and falling of the water give rise to *seiche currents* which in narrow straits may attain a fairly high velocity. It is well known that the wind tends to drag the surface layers of a water body along with it, thus producing within a very short time a distinctly noticeable *wind current*, or "wind drift" as it is more often called. The great systems of prevailing winds combined with the modifying effects of the earth's rotation, the forms of land masses, and other factors, have generated permanent systems of currents in the principal oceans. These are developed on a gigantic scale and are commonly distinguished from the local currents produced by wind action alone.

Since currents of this type must be characteristic of any rotating planet which possesses an atmosphere and oceans, we may refer to them as *planetary currents.* Barometric pressures being greater in one place than in another, water may, under favorable conditions, flow from the region of high pressure toward that of low, as more or less distinct *pressure currents.* If one portion of the ocean is more highly heated than another, the difference in density between the lighter warm waters and the heavier cold waters will give rise to *convection currents* by means of which the waters will endeavor to re-establish a condition of equilibrium. In much the same manner oceanic waters, which are more saline and therefore heavier than adjacent waters, will produce exchange currents with the lighter, less saline waters. We may for convenience call movements of this origin *salinity currents.* When rivers enter the sea their currents are progressively checked as they proceed farther and farther into the quieter water; but for some distance out from shore there may often be recognized very distinct *river currents.* The dynamic force of these outflowing streams causes bottom currents which move landward into the river mouths, and which have been called *reaction currents.* A current of any origin may be accompanied by lateral whirls or eddies, and these may be of sufficient diameter to give *eddy currents* of considerable importance. Whenever any one of the above types of currents impinges upon a coast, there results a piling up of the water with the consequent establishment of an hydraulic gradient. Water will flow from the higher to the lower level, and the resulting currents will here be spoken of as *hydraulic currents* ("polarization currents" of Cornish[1]). A valuable discussion of the theory of some of the above mentioned types of currents will be found in a series of papers by V. W. Ekman[2] published in the "Annalen der Hydrographie und Maritimen Meteorologie" in 1906.

We will now consider the essential characters of the several types of currents in the order named above, except that it will be more convenient to treat the varieties of hydraulic currents in connection with the original currents which give rise to them. We shall purposely omit consideration of currents which are as yet but little known, such as the pulsating currents discovered off the Norwegian coast and described by Helland-Hansen[3], and the deep vortices in the Norwegian sea described by Helland-

Hansen and Nansen[4]. We must not forget, however, that some of the movements thus omitted may ultimately prove of importance to the student of shoreline topography, for, in the language of the author last named, " the sea in motion is a far more complex thing than has hitherto been supposed," and, " there must be many forms of motion of great and far-reaching importance, though hitherto hardly known at all."

Wave Currents. — It has already been shown that in normal oscillatory waves the water, from the surface downward, moves forward under the crest of each wave and backward under the trough. In shallow water this alternating current movement is accomplished without any accompanying vertical motion in that part of the water next to the bottom. The forward and the backward currents are approximately equal in duration, and on a level sea-bottom should nearly compensate each other. There will be a slight advantage in favor of the forward current, due to the slight progressive motion of the water particles in the direction of wave propagation which Stokes has shown to exist in oscillatory waves[5]. On shallow bottoms this would result in a slow advance of movable débris in the direction of wave propagation.

Since the depth of wave action depends mainly upon the length of the waves, it is evident that the long groundswells which come from distant storms will affect the bottom waters more than will shorter storm waves developed near the coast. Storm waves may move landward or oceanward, according to the wind direction; but the swells always move landward. Hence the waves which most affect the bottom come on-shore; and the advantage resulting from the slight excess of the forward component of wave motion will be exerted mainly in a landward direction. On a level sea-bottom this would mean a transportation of movable débris prevailingly in a landward direction.

The advantage just referred to is often more than offset by the seaward slope of the bottom. If particles of débris are given a certain impulse up the incline, against the pull of gravity, they will travel a comparatively short distance; if given a nearly equal impulse down the incline, in the direction of the pull of gravity, they will move a distinctly longer distance. Thus an alternating current with a slight excess of the shoreward component may cause a *seaward* transportation of débris on the

ordinary offshore slope. Many authors, as for example Cornaglia[6], do not assign sufficient importance to the effects of gravity on a steep slope, the undertow, and other seaward-acting components to be discussed later; but consider that the normal consequence of wave action on the bottom is ordinarily to produce a landward advance of débris of proper size and specific gravity.

Where the water is so shallow in comparison with the wave length that there is produced a steepening of the wave front, another element is introduced. As Cornish[7] has pointed out, under these circumstances the forward motion is quick and short, the backward motion slower and of longer duration. This means that the shoreward component of such waves is much the more effective in moving coarser débris, since a shorter lived current of high velocity will transport material which is too large to be moved at all by the longer enduring but weaker seaward current. Sand and silt, on the contrary, will readily be moved a nearly equal distance in both directions, or on a sloping beach the seaward movement may predominate, as already shown. From this it follows that the same waves may drive pebbles and cobblestones toward the beach and finer débris toward deep water, at one and the same moment. Or, as Cornish has well expressed it, "suitable oscillation on a seaward slope will set shingle travelling shoreward, and sand simultaneously travelling seaward "[8].

A further reason for the landward progress of coarse débris during wave action is elaborated by Cornish in his book on "Waves of the Sea and Other Water Waves "[9]. He shows that the forward current begins just as the vertical component of wave motion is raising coarse material from the bottom, with the result that this material is readily carried forward while in suspension; whereas the backward current sets in while the water particles are descending in their orbits and are therefore depositing coarse material upon the bottom where it is less effectively moved. This argument loses much of its force because Cornish takes no account of the fact that on a smooth bottom the oscillatory motion of the water particles is backward and forward in a horizontal plane, the vertical currents upon which the validity of his theory depends being absent. Immediately above the bottom the vertical element of the oscillation begins to appear, and material carried upward a sufficient

PLATE X.

Photo by Spurr and Son.

Cobblestones cast upon the beach from deep water through the combined effect of wave action and the buoyancy of attached seaweeds.

distance by eddies due to inequalities of the bottom might be somewhat affected in the manner described.

If we turn our attention for a moment to the action of normal waves of translation, we have to note that the currents which they produce constitute essentially one intermittent current acting in a uniform direction. The water particles, from the surface to the bottom, move forward and then stop, the process being repeated with the passing of every such wave. Accordingly the débris on the bottom is always urged forward; and since these waves usually come on-shore, they give rise to a landward progress of all movable material, both fine and coarse. Russell attributes the shoreward transportation of shingle and wreck to the action of waves of translation[10]. It appears certain that either waves of translation or oscillatory waves may, under proper conditions, effect a very remarkable transport of débris toward the land; for Murray[11] has shown that shingle and chalk ballast dropped into the sea off Sunderland at a distance of 7 to 10 miles from land, where the water is from 10 to 20 fathoms deep, are thrown on shore by storm waves; and Gaillard quotes Robinson as authority for the statement that at Madras, during a violent storm, a quantity of pig lead, which proved to have come from a vessel wrecked more than a mile offshore[12], was cast upon the beach. The landward transport of large cobblestones from deep water far offshore is often effected " not by the simple impulse of the currents or storm waves, but by such action combined with the buoyancy given to the stones by the growth of seaweed attached to them " (Plate X), as was pointed out by Kinahan[13] many years ago. Shaler[14] believes that some shingle beaches receive their entire supply of material in this manner. In the waves produced experimentally by Caligny, which combined a translatory movement of the water with an oscillatory movement, particles on a level bottom were transported in a direction opposite to that of the wave propagation[15]; but we have no sufficient evidence that waves of this type are common in nature.

When a wave breaks at the foot of a beach slope, the water which is driven up the slope, forming the *swash*, carries material landward, while the *backwash* tends to transport it seaward again. The landward component of this alternating current is as a whole the stronger, because the return current suffers a loss of ve-

locity due to the friction which acts continuously, and a loss of volume due to percolation of the water into the crevices between the sand and coarser material composing the beach. Where the beach slope is very steep, however, the seaward current may be the more effective of the two because it works with gravity, while the landward current must propel the débris up the slope against the pull of gravity. It should be noted that while the swash of the wave, advancing up the beach slope, may retain something of the forward component of the true oscillatory motion belonging to the wave at the moment of breaking, the backwash is really an hydraulic current containing no element of true wave motion.

Beach Drifting. — If waves break obliquely upon a beach, there results a very important longshore transportation of débris on the beach slope itself. To distinguish this phase of shore activity

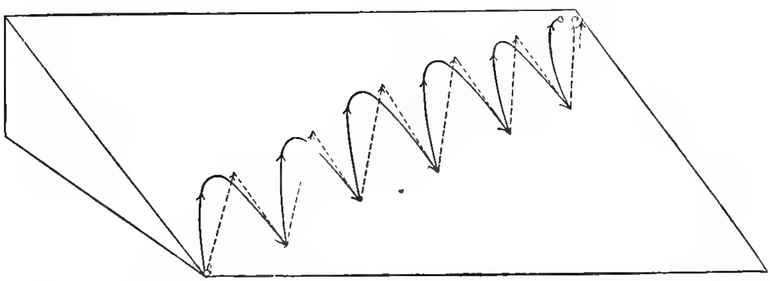

Fig. 13. — Section of beach slope showing by dotted lines the so-called zig-zag path of débris particles during beach drifting and by solid lines the parabolic paths actually followed.

from the longshore transportation effected by currents in the water just outside the beach, I propose to call it *beach drifting* ("Strandvertriftung" of Krümmel, "Küstenversetzung" of Philippson). Conformable to this usage, the longshore transportation which takes place in the shallow water seaward from the beach, often called "longshore drift," will be termed *longshore drifting.* The terms *beach drift* and *longshore drift* will then be restricted to the material transported by these processes, both being included under the broader term *shore drift.* In the case of beach drifting, the swash of the wave advances obliquely up the slope, continuing the direction of advance of the wave; but the backwash,

PLATE XI.

Surf breaking on the shore of Cape Canaveral, Florida. Nearest the observer is seen the swash of a broken wave ascending the beach slope.

being under the control of gravity, tends to return directly down the steepest slope. As a matter of fact, the control of gravity replaces the oscillatory movement of the water gradually instead of abruptly, with the result that the water does not advance in an oblique straight line to the top of the beach slope and then descend in a straight line at right angles to the shore, describing the zig-zag path shown by the dotted lines in Figure 13, but rather describes a series of parabolic curves as shown by the solid lines. It follows that a pebble under the influence of such wave action does not strictly speaking, pursue a zig-zag course along the beach as is usually stated, but rather a course represented by parallel parabolas. Palmer[16] was the first, so far as I am aware, to call attention to the importance of this phase of wave activity in causing a longshore transportation of débris; but his figure illustrating the process of beach drifting incorrectly represents a zig-zag path for the transported material, and contains a still more serious error in that it represents the swash as carrying both large and small particles an equal distance up the beach slope, although he recognized that small particles were carried farther down the slope by the returning backwash. Figure 14 illustrates the fact that small particles describe bigger parabolas than larger débris, and therefore progress along the beach with greater rapidity.

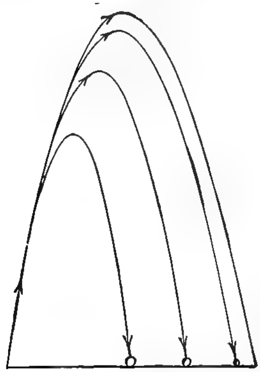

Fig. 14. — Parabolic paths of large and small particles of débris subject to beach drifting.

The positions of the parabolic paths taken by the particles in beach drifting usually depend upon the combined action of more than one set of waves, as when the surf and a superposed set of wind waves strike the beach at different angles. Even if the surf breaks parallel to the beach there will be some beach drifting if the wind waves arrive at an oblique angle; but, as shown by Figure 15, the angle of advance of the water up the slope will not be as oblique as if determined by the wind waves alone, since the path actually taken is the resultant of the impulses given by both waves. A longshore tidal current, or any other current parallel

to and near the shore, may combine with waves which break directly on shore to give a very pronounced beach drifting in the direction of the current. With a longshore current moving in a direction opposed to that of the oblique waves, sand may travel

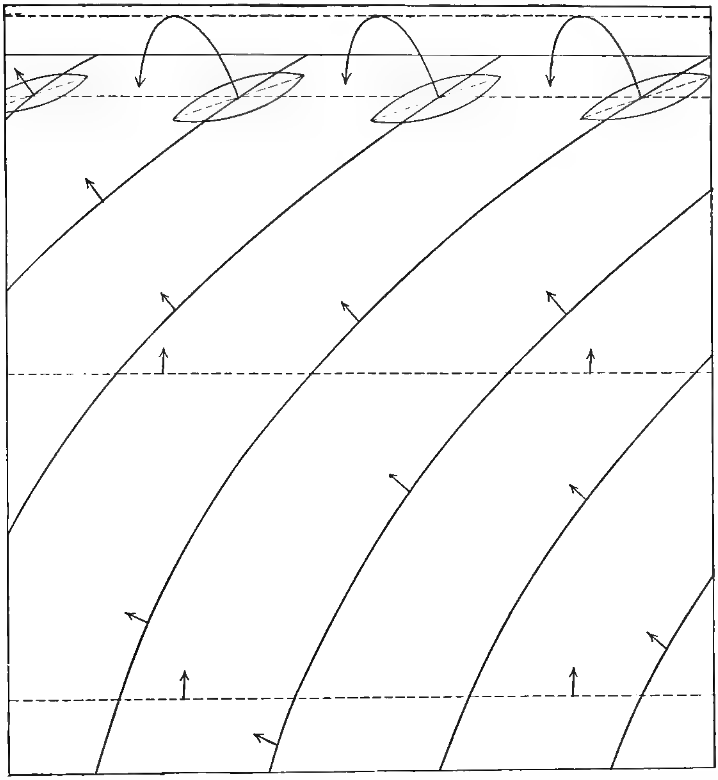

FIG. 15. — Parabolic paths followed by débris particles impelled by the combined action of onshore swells (broken lines) and oblique wind waves (solid lines). After Krümmel.

with the current and coarser material with the beach drift, as was fully recognized by Owens and Case[17].

The direction of beach drifting will depend upon many factors. Among these may be noted the direction from which the groundswells approach the shore, in those cases where they are not sufficiently refracted to strike the beach at right angles, and in which wind waves are on the whole less powerful in determin-

PLATE XII.

Winthrop Great Head, a wave-cliffed drumlin near Boston, Massachusetts. Débris cut from the drumlin by waves has been transported by beach drifting and built into bars, one of which is seen in the foreground.

ing the movement of shore débris. Another important factor is the direction of the prevailing winds, or of the dominant storm winds, in case these develop waves of considerable power, and the groundswells are weak, or do not approach the shore obliquely. The direction of the greatest stretch of open water is likewise important, since weak winds blowing over a long stretch of water may develop larger waves than strong winds which cross a limited water area. A good example of the effect of "length of fetch" is found in the beach drifting along the sandspit which encloses Toronto Harbor on Lake Ontario. Here the movement of the beach material is westward against the prevailing westerly winds, because the greatest stretch of water over which westerly winds can blow is 40 miles, whereas easterly winds cross 180 miles of the open lake surface[18]. Failure to recognize the important relation of beach drifting to the direction of greatest expanse of open water has led many authors to unsound conclusions, a typical example being Haupt's arguments against the efficiency of beach drifting along the New Jersey and other shores based on the assumption that if there were any effective beach drifting it would have to move with the prevailing winds[19].

Haupt cites the well known fact that on the Great Lakes material may be drifted in opposite directions from some point near the middle of one side of a lake, and concludes this is sufficient proof that wind waves cannot be responsible for the movement. An inspection of Figure 16 will suffice to show that this conclusion is not justified. Since the dominant waves (shown by heavy lines in the figure) depend upon length of fetch as well as upon intensity and duration of the wind, it is evident that beach drifting between *a* and *c* will be southward; because the winds from the northeast, blowing across a broad stretch of open water, will generate more powerful waves than the winds from the southeast which traverse a shorter stretch of water, or the much more important prevailing winds from the southwest which blow directly off the land. Beach drifting from *a* to *c* is thus opposed to the direction of the prevailing winds. For similar reasons the material north of *a* is drifted in the opposite direction, toward *b*; and on the east side of the lake material is drifted in opposite directions from *d*. The expectable directions of beach drifting derived theoretically in the accompanying diagrams (Fig. 16) appear to correspond with the actual directions re-

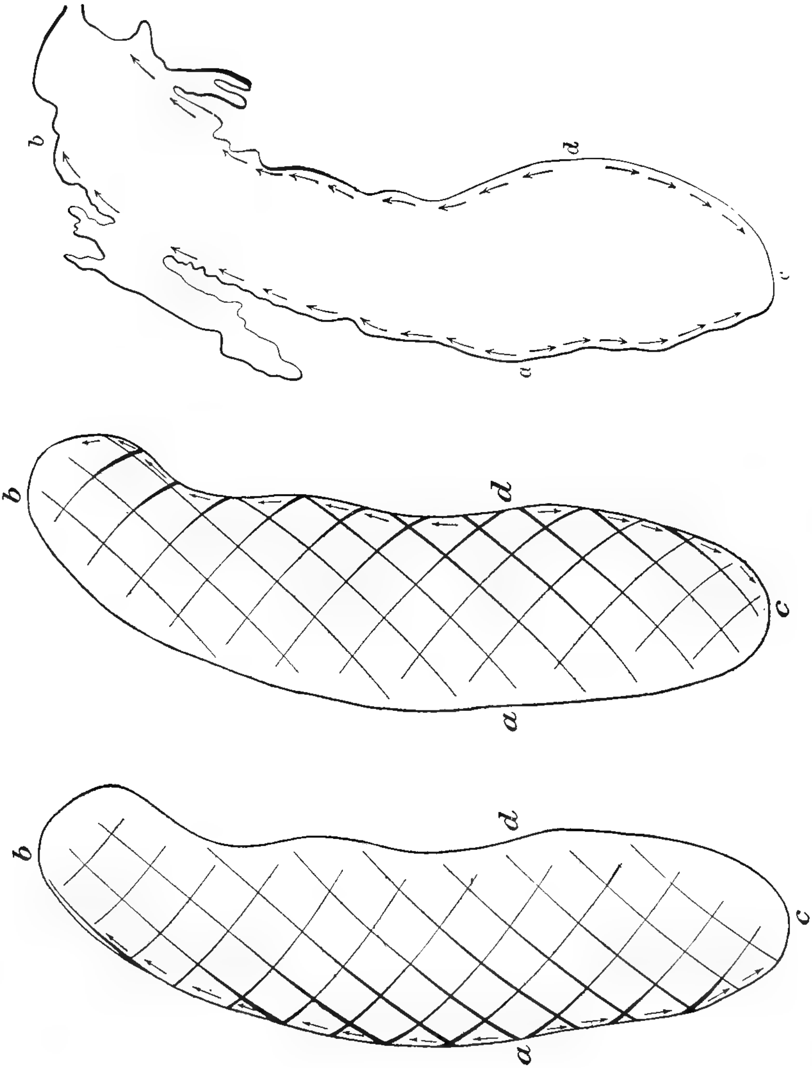

Fig. 16. — Diagram to illustrate relation of beach drifting to wind directions in an ideal case and in the case of Lake Michigan. The first two figures show the relative intensities of oblique wave action and the direction of beach drifting on the western and eastern sides respectively of an ideal lake with winds blowing from all quarters. The third figure shows reported direction of beach drifting along the shores of Lake Michigan.

ported for Lake Michigan[20], a lake somewhat similar in form to the ideal lake of the figure and similarly situated with reference to the prevailing winds. On Lakes Erie and Ontario Wilson[21] finds a similar relation between direction of beach drifting and length of fetch of open water. A proper appreciation of this simple principle will enable one to understand many disputed

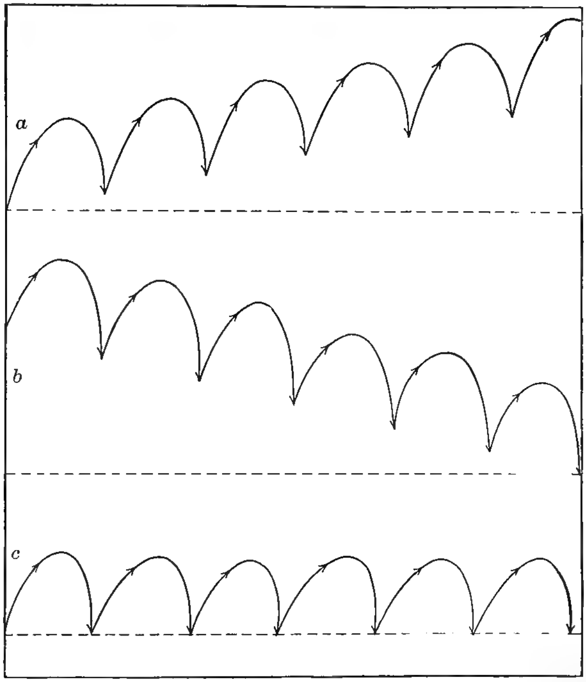

Fig. 17. — Parabolic paths of débris particles subject to beach drifting on (*a*) a prograding beach, (*b*) a retrograding beach, and (*c*) a graded beach.

points regarding the movement of shore débris on irregular sea coasts.

Beach drifting may occur on a prograding beach, on a retrograding beach, or on a beach which is at grade, i.e., one which is neither losing nor gaining material. On a prograding beach the relation of slope to volume and velocity of the alternating currents is such that particles advance farther than they retreat, and the ideal path of a single particle is represented by Figure 17*a*. The paths which particles on a retrograding

PLATE XIII.

Lowestoft Ness on the east coast of England. A beach plain built in front of the former marine cliff (in foreground) largely by beach drifting. The fisherfolk spread their nets to dry on the level surface of the beach plain.

beach and on a graded beach would tend to take are shown at *b* and *c* of the same figure.

The so-called "zigzag" progression of particles of débris is commonly treated in connection with beach drifting only, as though this type of movement were restricted to the zone of breaking waves on the shore. Cornaglia[2] was right, however, in ascribing such a movement to débris on the sloping bottom seaward from the beach during the passage of unbroken oscillatory waves in a direction oblique to the slope. On the bottom the motion of the water particles, as we have already seen, tends to be in a straight line, back and forth. With waves oblique to the slope this bottom movement (called "flutto di fondo" by Cornaglia) would carry the material obliquely up and down the slope over the same path, with a general advance or retreat in the same straight line when onshore or offshore components prevailed, were it not for the effect of gravity. Under the influence of this force both water particles and transported débris tend to return more directly down the slope after each forward oscillation, with the result that a progressive motion, parallel to the shore, is added to the back and forth movement. Or, in common parlance, the particles "pursue a zigzag path" (more properly a series of parabolic curves) on the sea-bottom, which results in longshore drifting of a type analogous to beach drifting.

We shall find in later chapters that many shore forms commonly attributed to tidal and other currents are more reasonably to be interpreted as the product of beach drifting. The water movements involved in beach drifting have a high velocity and hence a great transporting power; and one may readily observe coarse débris carried along the coast by their force when tidal and other currents are too weak to move anything but the finest sands. Shaler has watched pebbles made from ordinary bricks move along the shore at the rate of more than half a mile a day under the influence of beach drifting, and Wheeler[23] observed half bricks carried 25 to 30 yards in from $1\frac{1}{2}$ to 2 hours by the same force.

Hydraulic Currents due to Waves.—Thus far we have been mainly concerned with those currents which are more or less directly involved in the normal oscillatory or translatory motions of the water particles in waves. We must now turn our attention to the hy-

draulic currents, which are the indirect product of wave action. It has already been shown that with every wave of translation there is a direct shoreward movement of the water, which is not compensated by a backward movement. Hence a series of such waves coming onshore tend to pile up the water above the normal level of the sea. Since oscillatory waves entering shallow water are partially transformed into waves of translation, they too must cause accumulation of wa er against the coast. Even were they not thus transformed, the slight excess of the shoreward component in waves of oscillation which has already been described would have a tendency in the same direction. Thus onshore waves raise the level of the sea along a coast upon which they break. An appreciable local rise in the sealevel due to this cause has been inferred by several writers[24] and has been demonstrated by the author at certain points along the Atlantic Coast.

It is clear that the water piled up against a shore in the manner just described must escape, thereby producing more or less continuous "hydraulic currents." If the escape is seaward, along the bottom, we have the current known as the *undertow;* if the escape is effected by currents moving along the shore away from the area of accumulation in either direction, we have a longshore current, sometimes called "longshore drift." The undertow may temporarily be checked under each wave crest, and may even have its direction momentarily reversed by the forward moving water of that part of the wave; but under the wave trough the undertow combines with the backward moving component of oscillatory waves to form a seaward bottom current of great strength. A marked development of the undertow is favored by oscillatory waves, for these disturb the bottom waters less than the surface; by a broad zone of waves striking a long stretch of the shore at right angles, since these conditions are unfavorable to the ready escape of the water as longshore currents; and by a steep offshore bottom and deep water close to shore, because the returning water is then enabled to pass quickly down beneath the disturbed surface and move seaward with little interruption. Longshore movement is favored by waves of translation, since waves of this class give a vigorous shoreward motion to all the water from the surface to the bottom; by an oblique angle of wave incidence, because

water propelled obliquely against a shore tends to produce a strong current in the general direction of the propulsive force; and by gradually shallowing water offshore, which favors the development of waves of translation and a shoreward movement of the water at all depths.

Work of Wave Currents. — We must conclude from what has been said in the preceding paragraphs that waves are profoundly important as agents of erosion and transportation, both on shores and shallow bottoms. It is not easy to understand the process of reasoning which led Lieutenant Davis to ignore the more important activities of wave currents in his memoir on the various currents of the ocean, and to conclude that "the most noted and interesting effect of waves is the ripple-mark"[25]. The careful reader of his memoir will discover that many of the phenomena ascribed by Davis to tidal action are more probably the effects of wave currents. In like manner Kinahan[26] reaches the conclusion that wind waves do very little permanent work. He ascribes to tidal action beach drifting and other phenomena undoubtedly produced by wave action.

It is also evident from the foregoing paragraphs that the action of wave currents upon débris varies greatly under different conditions. On a flat bottom oscillatory waves will move débris prevailingly shoreward; but if the slope be steep enough, the same waves may cause material to migrate seaward; or coarse débris may be propelled shoreward and fine débris seaward. If the waves belong to the class of true waves of translation, the débris will be transported landward, even on a sloping bottom. Waves breaking on the beach drive material up the slope until continued accumulation makes the slope so steep that the backwash returns all material to the breaker zone. If the beach slope is too steep for a given set of waves, the backwash will return more material than was brought by the forward rushing current, and the beach will suffer erosion. Beach drifting will vary in direction and amount with changes in the direction and size of the waves. The seaward component of wave motion may be effectively supplemented by the undertow. If the undertow is strong it may prevail over the landward component of wave motion, and cause the bottom débris to move continuously seaward; but if the waters piling up against a coast escape laterally as longshore currents, the débris may

first move landward, and then suffer some longshore transportation under the influence of these currents. Since the several types of wave currents vary in strength with the outline of the shore, the angle of offshore slope, the angle at which the waves approach the shore, the size of the waves, and the kind of waves, it is manifest that the analysis of wave action upon shore débris is no simple matter. This conclusion is amply justified by the experience of those engineers who have studied the effects of waves on natural shores and artificial structures. Gaillard[27] expresses the general opinion of the profession when he says " In scarcely any branch of engineering are the forces developed and the methods and directions of their application more variable than in the case of wave action." Before pursuing this point further, let us proceed with our inquiry into the behavior of other types of currents.

Tidal Currents. — The tides may best be considered as great waves which combine some of the features of both oscillatory

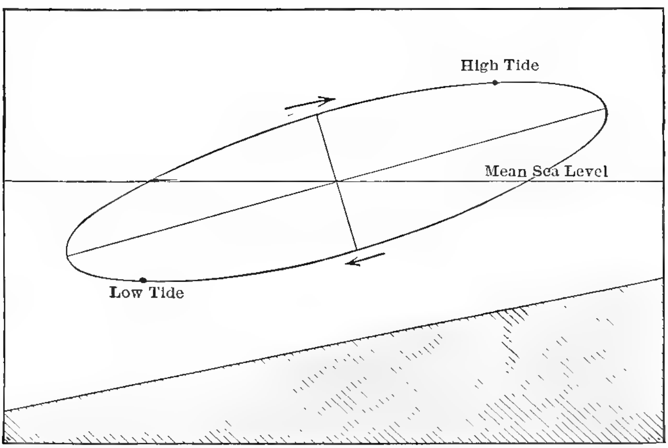

Fig. 18. — Elliptical orbit of water particle during passage of the tide wave over a sloping sea-bottom.

waves and waves of translation[28]. They resemble oscillatory waves in having an orbital motion of the water particles, the orbit becoming a very much flattened ellipse in the shallowing water, with its long axis rising toward the land (Fig. 18). As will appear from the figure, there is a shoreward movement

of the water particles until near the time of high tide, after which a seaward movement takes place. These orbital movements of the water constitute tidal currents. Immediately at the shore the landward or "flood current" may continue to flow until the very moment of high tide. In the open sea, or in the case of tides passing a headland projecting far out to sea, the orbital path would not be distorted as in Figure 18, but would be more nearly circular; hence it is clear that the landward movement would persist for a long time after high tide, just as the forward motion of the water particles in an oscillatory wave continues after the wave crest has passed (Fig. 1).

The great importance of these tidal currents may readily be appreciated from a consideration of their velocities. Krümmel has shown that in water 30 meters deep a tidal rise of 3 meters should result in currents having a velocity of 1.7 knots per hour; and with a rise of 4.5 to 6 meters the currents should attain a velocity of over 3 knots per hour. The observed velocities are in agreement with the theoretical deductions. According to Wheeler tidal currents in the English Channel between Scilly and Hastings have a velocity of 2 miles * an hour; in the northern part of the Wash, 4 miles; and off the island of Ushant, France, 6 to 7 knots per hour[29]. In St. Malo Bay where there is a rise of 10 to 12 meters and a water depth of 30 meters, the velocity of tidal currents is from 5.1 to 6.7 sea miles per hour[30]. At Hell Gate in New York Harbor the currents attain a velocity of 4.8 knots per hour[31] while Bailey reports a current of "not less than 8 knots" through the Petite Passage southwest of Digby Gut, Nova Scotia[32]. Stevenson gives the velocities of a dozen tidal currents which vary from a minimum of 5.75 to a maximum of 12.20 statute miles per hour[33]. Sollas states that the tides in the Severn estuary have a velocity of from 6 to 12 miles an hour[34], while Krümmel cites velocities of 8 to 10 knots between the Orkney and Shetland Islands, 11 knots in the dreaded "Roost" of Pentland Skerries, and 11½ knots in the Gulf of Hangchau[35].

* It has seemed wisest to give velocities in the units originally employed by the various authorities, as any attempt to convert the expressions into a standard unit of measurement would in some cases introduce a misleading appearance of accuracy if fractional parts of the unit were employed, and in other cases would introduce large errors if the fractions were ignored.

The transporting and eroding power of such currents is enormous. A velocity of but .4 knot per hour will drive ordinary sand along the bottom, while fine gravel will be moved if the velocity rises to 1 knot; shingle about an inch in diameter is moved at 2.5 knots; and angular stones about one and one-half inches in diameter, at 3.5 knots[36]. Inasmuch as tidal currents continue for many miles in the same direction, it is evident that they must play a very important rôle in the transportation of shore débris and in submarine denudation whenever the velocity approaches the higher figures mentioned above.

G. H. Kinahan describes a number of beaches and submarine banks on the coast of southeast Ireland which he believes were formed mainly by tidal currents[37]. H. C. Kinahan states that sands and gravels in "Beaufort's Dyke" off the coast of the Mull of Galloway are moved back and forth by currents generated by the combined action of tides and waves at a depth of 720 to 860 feet[38]. Along the deeper middle portion of Long Island Sound the mean velocity of the tidal inflow is nearly 1 meter per second and of outflow slightly less, or high enough to transport coarse gravel. Dana shows that wherever there is any narrowing of the Sound by shoals or islands there is an increase in depth, and he attributes this to increased erosive force of currents at these points. He finds such effects to a depth of 330 feet[39]. In a paper discussing "Erosion durch Gezeitenströme" Krümmel expresses the opinion that this agency is responsible for the fact that whereas the floor of the Bay of Fundy usually has a depth of from 50 to 70 meters or less, depths of from 100 to 110 meters occur where the tidal currents are restricted by the narrows at Cape d'Or and Parrsboro[40]. Reade ascribes to tidal scour the formation of trenches between islands off the coast of Scotland having depths of nearly 800 feet[41]; but the possibility that these trenches represent submerged subaerial valleys should not be overlooked. The strong tidal currents of the Severn sweep along great masses of boulders thereby deepening the channel, according to Sollas; and Richardson attributes the deep water known as the "shoots" to this erosive action[42]. Sections taken along the deep-water channel of the Hooghly River in 1813 and 1836 showed that between those years tidal currents had scoured out the silt of the river bed to a depth of 52 feet, forming a "scour hole"

20,000 feet long at the top and 9000 feet long at the bottom[43]. Helland-Hansen[44] has shown that marked tidal currents exist at the bottom of fairly deep oceanic waters. On the Michael Sars Expedition, which made the first measurements of such currents in deep water, he found a true tidal movement of .27 meter per second (more than .5 knot per hour) at a depth of 732 meters, or 2400 feet, south of the Azores. Krümmel[45] admits the efficiency of tidal currents in sweeping rocky ridges free of mud at a depth of 6500 feet or more. When it is remembered that a current of .20 meter per second or .4 knot per hour will transport ordinary sand, it is clear that tidal currents may transport coast débris to great depths and under favorable conditions may even effect some erosion far below the surface of the ocean. Gardiner[46] has gone so far as to attribute the submarine plateau of the Maldives to the action of planetary and tidal currents in cutting down a land area to a depth of 1140 feet below sealevel; but while the theoretical possibility of such erosion must be admitted, the evidence on which Gardiner bases his conclusion in the Maldive case is not convincing.

Tidal currents do not always, or even commonly, act in a direction normal to the shoreline. Along the sides of a bay or headland whose axis is in the line of tidal advance, the current may be parallel to the shore. Shoreline irregularities will deflect the tidal waters, giving longshore currents in all possible directions. These longshore currents are commonly much swifter than those movements which take place normal to the beach. On an open coast, exposed to the direct advance of the tidal wave, the onshore and offshore movements of the water are very weak, and can accomplish very little geological work, as can readily be verified by the observer on such a coast during a calm day. On the other hand, longshore currents close to the land are very effective geological agents, since they remove the débris produced by wave erosion and brought to the sea by rivers, transport it to distant localities, and often deposit much of it in deep water. They may even produce profound changes along the shores by direct erosion, as in the case of the violent currents associated with the bore in the estuary of the Amazon, the effects of which have been well described by Branner[47]. The currents which pass up and down a bay or estuary are here considered longshore currents; for while they may be

PLATE XIV.

Photo by E. M. Kindle.

Giant sand ripples produced by strong ebb tide current in the Avon River estuary near Windsor, Nova Scotia. The sand is moved rapidly in the direction of the tidal current.

normal to the general trend of the outer coast, they are in general parallel to the immediately adjacent shores.

It is a well known fact that a narrowing bay compresses a tidal wave into smaller space and constrains it to rise higher. Thus we get the remarkable tidal rise at the head of the Bay of Fundy and in the River Severn. It is likewise true that when the energy of the tidal wave is transmitted to the smaller volume of water in front, the effect on the latter is correspondingly great. The smaller volume of water develops a swifter current and piles up higher against the coast. If the form of the coast prevents the escape of the accumulated waters laterally, they will continue to rise until the head counterbalances the momentum of the advancing current. On the other hand, if a large bay is separated from the open ocean by a narrow inlet, practically no true tidal motion takes place within the bay. The tidal wave is scarcely transmitted through the narrow channel, and the water within the bay rises because of the hydraulic head resulting from the accumulation of water against the coast outside. Such currents as result from the rise and fall of the water within the bay are really hydraulic currents, and are not parts of any true oscillatory movement of the water.

In bays and sounds the swiftest tidal currents follow the deepest channels, and are therefore not as directly effective in shore processes as when they impinge upon an exposed portion of the coast. Even here, however, they have an indirect effect of no mean importance; for they remove vast quantities of débris, which was originally eroded from the land by wave action or carried to the sea by rivers, and then transported by longshore currents of different types until brought within the influence of the inflowing or outflowing tidal current. The inflowing tide sweeps the finer material far up the bay where it is deposited in mud flats and tidal marshes, which are often reclaimed for cultivation by the process known among the English as "warping"[48], while the coarser sand is moved landward a much shorter distance, often forming bars along the channels. The outflowing current carries the material it receives out to sea, and shifts the bars in the same direction. Because of the river water usually poured into a bay the ebb current predominates over the flood, and the direction of débris migration is prevailingly seaward. The net result of this current action, therefore, is to

PLATE XV.

Salt marsh near Green Harbor, Massachusetts. Deposition from tidal currents and the growth of marsh vegetation have filled all the area except the channels of tidal creeks. The houses stand on a bar separating the marsh from the ocean.

favor wave erosion by removing débris to deep water, thus keeping the shores better exposed to renewed attacks.

Deposition by Tidal Currents. — The great importance of incoming tidal currents in bringing about local deposition at the heads of bays and in the quiet waters of harbors justifies further consideration of this point. There is a tendency to ascribe to the deposited material a fluvial origin, and many have argued that only the rivers entering the bay are capable of bringing so much fine sediment to the place of deposition[49]. But Skertchley[50] has shown that the rapid silting up of the Wash in eastern England, by which a breadth of three miles has been added to the land in some places since the Roman occupation, is accomplished by the sea and not by rivers. Deposition occurs mainly at the slack of high water, although a little of the material settles in sheltered places during the ebb. Crosby[51] ascribes the deposition of silt in Boston Harbor to the action of incoming tidal currents, and shows that the Mystic River, which enters the harbor, has effected scarcely any deposition in the lakes through which it flows during the same period of time in which a maximum of 25 feet of the silt has accumulated in the harbor. Mitchell[52] is of the same opinion regarding detritus underlying salt marshes on other parts of the New England coast. The extensive deposits of red silt at the head of the Bay of Fundy are largely due to the action of the strong flood tide which carries in the material eroded from the shores of the bay[53]. In his excellent study of the Severn estuary Sollas[54] has demonstrated that the flood tide not only brings in large quantities of silt but also innumerable remains of marine organisms which are deposited with the silt in the upper reaches of the main estuary and in the tributary estuaries. According to Browne[55] this deposition takes place not only at the slack water of high tide, but during two-thirds of the ebb tide, since he found that in the Avon below Bristol the silt-laden lower waters remain stagnant long after the surface waters have begun to ebb.

It should be appreciated, however, that even where extensive deposits of silt are laid down by tidal action at the heads of bays, these same tides transport much material far out to sea, where it comes to rest in deep water. On a subsiding coast the amount of material deposited at the bay head may exceed that carried seaward; and if the subsidence gives place to stability

this excess of deposition may continue for a time, until the heads of the drowned valleys are well silted up. In time, a condition of approximate equilibrium will be approached, when the swifter tidal currents of the narrowed channels will erode about as much material as they deposit. From that time onward the material brought out by rivers and eroded from the shores by waves will add little if anything to the extent of the tidal deposits. The tidal currents charged with sediment will sweep up and down the bay, depositing in one place and eroding in another, depositing at slack water and eroding at times of swiftest flow; but each retreating tide, re-enforced by the outflowing river water, will remove from the bay an amount of material equivalent to that brought in by various agencies. According to Sollas the Severn estuary has reached this nicely balanced condition, in which "the accumulation is always being diminished by withdrawals seaward, and as constantly renewed by fresh accessions provided by the denudation of the land"[56]. The Wash appears not to have reached this stage of equilibrium, for, as described by Skertchley[57] and observed by the present writer, the deposition far exceeds the removal of material and the land gains upon the sea. Along the head of the Wash the average rate of gain was 7.29 feet per year from the second to the seventeenth centuries, 48.65 feet per year during the eighteenth century, and 31.68 feet per year during the nineteenth century[58]. It would seem reasonable to suppose that with a gradual decrease in the supply of sediment furnished to the tidal currents, an area of tidal deposits might pass beyond the stage of equilibrium and enter a stage in which more material was eroded from the region than was returned by the incoming tide. It is possible that the head of the Bay of Fundy has entered this last stage, for in several localities visited by me no appreciable accumulation had taken place since the last dykes were built, and in one locality considerable erosion had evidently occurred, uncovering the ancient forest described by Dawson in his Acadian Geology[59]. Perhaps an excess of erosion over deposition is also responsible for the abandonment of certain tide marsh areas and for the increased difficulty of maintaining the dykes; facts for which Dawson suggested a change in the direction of tidal currents as one of several possible explanations[60]. A careful examination of older and later surveys of regions about the head of the bay might

possibly determine the validity of the explanation here tentatively suggested.

In the case of a bay which receives little or no river water the tidal régime may be such that the flood currents prevail over the ebb at all times. Deposition will then exceed erosion, not merely until the regions adjacent to the main channels are silted up, but the channels may themselves be blocked and the tides completely excluded from the former bay by their own deposits. This fact led Browne to the conclusion that tidal deposition always exceeds tidal erosion, and that therefore when no river water flows through a tidal creek or bay to keep the channels scoured out, such an area must in time silt up entirely[61]. The arguments used to support his conclusion are not convincing, and it is probable that whether or not deposition exceeds erosion in such a bay or creek will depend on the nature of the tidal wave entering the depression and the nature of the currents to which it gives rise. Both vary greatly under different conditions, and there is no theoretical reason, at least, why the tidal régime may not be such as to favor erosion more than deposition in some cases, deposition more than erosion in others.

Movement of Débris by Tidal Currents. — The seaward journey of sediment held in suspension by tidal currents in an estuary or tidal river is far from simple. Even if we leave out of consideration the shorter or longer halts made by a given particle, and imagine it to be continually in transit, the ebbing and flowing tides carry it back and forth over the same ground many times, greatly prolonging its journey. Because the "land water" poured into the estuary by rivers causes the ebb tide to predominate over the flood by a greater or less amount, the particle is carried seaward by each ebb a little farther than the following flood carries it back; and so it gradually makes its way farther and farther toward its final resting place in deep water. An exception to this occurs temporarily in some estuaries where the resultant is landward while the spring tides are strengthening; but this temporary upstream progress gives place to a more pronounced seaward advance after spring tides are past. Numerous experiments with nearly submerged floats have demonstrated the predominance of the seaward component in such tidal oscillations. Figure 19 shows the course taken by such a float in New York Harbor, where the Metropolitan Sewerage Commission has

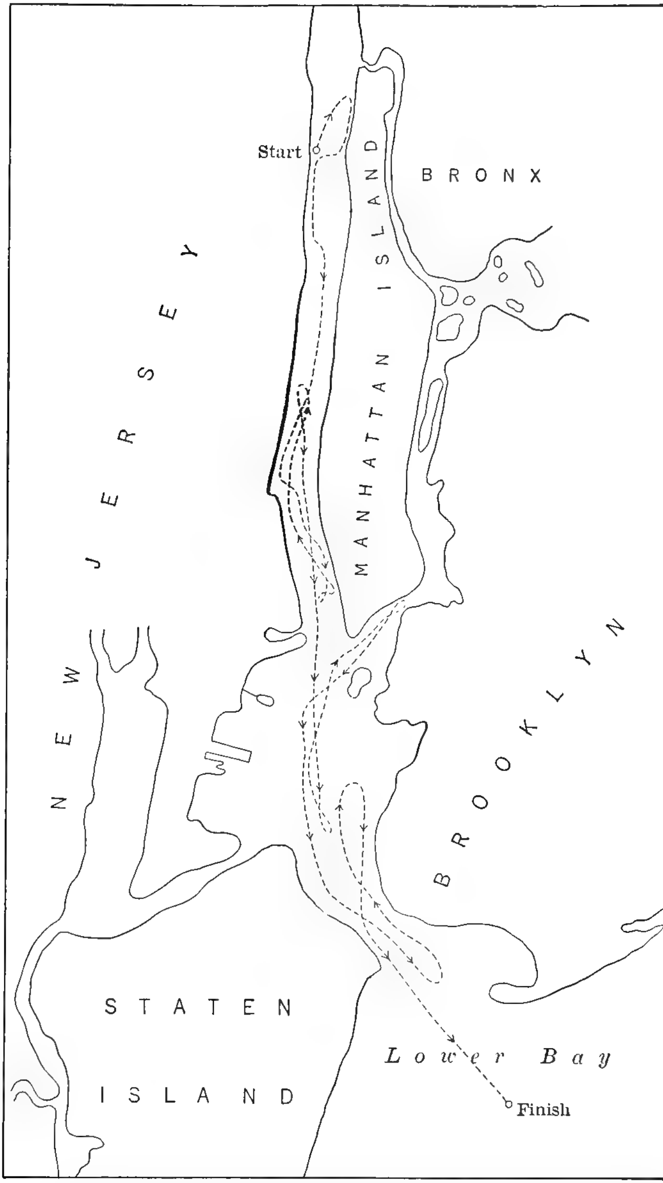

FIG. 19. — Course followed by a nearly submerged float under the influence of tidal currents in New York Harbor. (After Parsons.)

studied the effect of tidal currents on the transportation of sewage. Figure 20 shows the theoretical path during successive tides, which a particle would take on this same journey, and it will be seen that the theoretical and actual paths agree closely. Both demonstrate the seaward migration of particles in suspension. The comparative volumes of the ebb and flood currents, responsible for this seaward migration, may be seen in the following table taken from a paper by Parsons[62]. The importance of river water in augmenting the ebb is clearly apparent from this table.

VOLUMES FLOWING ON EBB AND FLOOD CURRENTS, HARBOR OF NEW YORK, IN MILLIONS OF CUBIC FEET

Part of harbor	Yearly means	
	Ebb	Flood
The Narrows	12,041	10,779
Hudson River, off the Battery	7,430	6,343
" " " 39th Street	6,990	5,903
" " " Fort Washington Point	6,230	5,143
" " " Tarrytown	3,980	2,893

In branches of the harbor where little land water enters, the difference between ebb and flood volumes is not so great. In other harbors where larger rivers than the Hudson enter, the difference must be much more marked. Experiments with floats in the Thames estuary show that the average seaward progression of a particle in suspension is $\frac{1}{3}$ mile a day[63].

One result of the back-and-forth journeying of each particle is the accumulation in estuaries and tidal rivers of a vastly greater amount of material than is daily contributed to their waters. "Thus in the waters of the Severn estuary there is a storage of suspended sediment, the accumulation of as many days, or weeks, or months as are occupied in its wanderings to and fro"[64].

There may be some question as to whether the coarser material on the bottom of estuaries and tidal rivers always has a tendency to move prevailingly seaward. Such material is certainly shifted back and forth by the ebb and flood currents, as shown

in the case of a sunken vessel at the mouth of the Gironde which had the sand scoured from about it by the ebb current and was completely buried again by the flood current[65]. Experiments with floats show the predominance of the seaward component of tidal oscillations at or near the surface of the water, but do not tell us about the movements in depth. It seems probable, however, that there is ordinarily a similar seaward tendency in the deeper waters also. The seaward component of the oscillation has a further advantage in that it works with gravity; for particles of sand will travel farther down the slope of the channel with the ebb than they will up the slope with a flood current of equal velocity and duration. But three factors at least tend to give an advantage to the flood current in certain cases. The waters of a bay are often much fresher than those of the ocean, because of dilution by

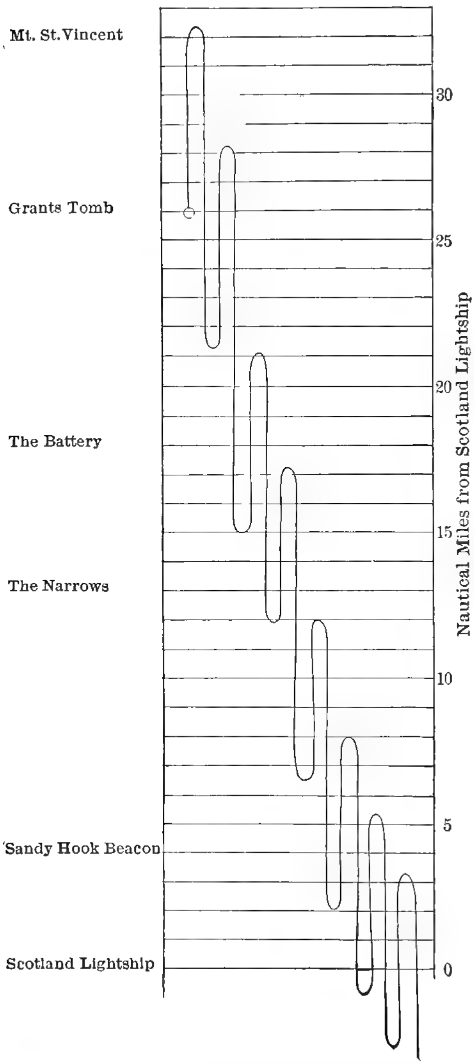

Fig. 20. — Theoretical course calculated by Parsons for the float whose actual course is shown in Figure 19.

rivers. Heavier salt water from the ocean may therefore push in along the bottom while the surface waters are still flowing seaward[66]. This action will be facilitated by differences in temperature, if the waters of the bay are warmer than those of the ocean[67]. Mitchell has shown that at the mouth of the Hudson River the bottom water moves landward with a velocity of .6 meter per second, or more than 1 knot per hour, while the surface waters are still ebbing[68]. During part of the ebb tide, therefore, transportation of bottom débris may be landward. Furthermore, since Browne[69] has shown that the bottom waters may also be stagnant during part of the ebb, it would seem possible to have a case in which the flood waters entered a bay along the bottom during the last of one ebb tide, then remained stagnant while the surface flowed back to the sea during much of the following ebb. If flood currents thus predominate on the bottom and ebb currents at the surface, coarser material will migrate landward while material in suspension is carried seaward. Mitchell was of the opinion that the flood does predominate in New York Harbor below a depth of 6 fathoms[70], and that it would cause the bar at the harbor mouth to advance up the channel, were this action not prevented by other currents[71]. As the coarser material of a landward moving deposit is ground to a finer size, however, it will rise in suspension and move seaward toward the ultimate goal of all land débris — quiet, deep water.

A second factor favoring the landward movement of bottom débris arises from the change in form which the tidal wave sometimes experiences when entering a bay or tidal river. The front of the wave becomes steeper than the back, and the flood current is much stronger than the ebb, the latter lasting a longer time. An extreme case of this inequality of current velocity is found in the tidal "bore" or "eager" which invades certain rivers, notably the Tsien-Tang-Kiang of China, where the front of the wave sometimes appears as a wall 25 feet high, and a million and a quarter tons of water may be carried by a given point in one minute[72]. The vigorous current of the "pororóca," or bore of the Amazon River, has already been mentioned[73]. Under such conditions bottom débris which is entirely too coarse to be affected by the longer continued but weaker ebb current may be carried forward by the flood. It would seem, therefore, that conditions may exist which compel a landward migration of

bottom débris under the influence of tidal currents, although this material when ground finer would move seaward under the same tidal régime. A sufficient body of observations is not yet available to enable one to determine how widespread these conditions may be.

Where beach deposits above mean sealevel are subject to transportation by tidal currents, either with or without the aid of wave currents, there may be a marked tendency for such shore débris to migrate in the direction of the flood current. This arises from the fact that where the tide flows freely, high water coincides more or less closely with the flood and low water with the ebb. Hence those beach deposits above mean sealevel will be moved by the flood current, but will not be reached by the ebb current. In bays and inlets this striking difference in the efficiency of flood and ebb currents in transporting beach material is less marked than opposite headlands, because near the bay heads flood begins when the tide is lower and ebb commences soon after high water is attained. It follows, therefore, that the débris which migrates along the shore from the headlands toward the bay heads under control of the dominant flood must come to rest where flood and ebb more nearly neutralize each other. This should give rise, in the absence of counteracting influences, to a tidal accumulation of débris in the heads of bays and inlets[74].

In a bay which has no strong tidal currents, the incoming flood may be incapable of stirring up any appreciable quantity of sediment, so that little material is carried to the bay heads for deposition. On the other hand, the ebb tide augmented by the outflow of river water may be sufficiently strong to carry into deeper water such sediment as is brought in by the rivers or supplied by wave erosion. Under these conditions tidal deposits at the bay heads will be conspicuous by their absence. The small amount of such deposits at the mouths of rivers entering Chesapeake Bay may perhaps be thus explained.

Along very irregular shores the comparative strength of flood and ebb currents can scarcely be predicted. Each area must be studied for itself. The positions of channels between islands and shoals, with reference to the direction of advance of the currents, may be such that some channels will have strong flood currents and very little ebb, while others will have vigorous

ebb currents and scarcely any movement during the flood. Bache[75] found the ebb currents near Sandy Hook much more powerful than the flood, and was indeed of the opinion that ebb currents are practically always far more important than the flood as eroding and transporting agents. The process of reasoning by which he reaches this conclusion does not seem convincing; and while the predominance of ebb currents in bays receiving upland waters may be admitted as a general rule, subject to certain exceptions, the relative strength of flood and ebb in the straits and other channels of an irregular coast must be more variable.

Hydraulic Currents Due to Tides.—The changes in surface level of the ocean resulting from tidal action inevitably cause the formation of various types of hydraulic currents, which we may now briefly consider. When the tide rises higher on one part of the coast than on another, any part of the water which does not participate fully in the tidal oscillation will flow from the higher toward the lower level under the influence of gravity. We may thus get hydraulic currents having the same periodicity as true tidal currents[76]. The waters piled up against a coast by a rising tide may escape to either side as longshore hydraulic currents; or if the waters are piled up at the head of a converging bay so that lateral escape is not possible, there may be developed an undertow which will give a seaward motion to the bottom waters before the direction of the surface current is reversed[77].

In the case of a bay separated by a narrow inlet from the open sea, the tide in the ocean rises so rapidly that enough water cannot pass through the inlet to keep the bay surface rising at the same rate. Later the tide in the ocean will fall more rapidly than the surface of the bay, because the outflowing water escapes through the narrow inlet so slowly. Consequently the ocean surface is highest part of the time, while the bay surface is highest at other times. These differences of levels, which may amount to a number of feet where the tidal range is large, give rise to hydraulic currents into and out of the bay. Such currents may have a very steep gradient and correspondingly high velocity, as in the case of those at the narrow entrance to St. John's Harbor, New Brunswick, where the average maximum head is nearly 10 feet, and a reversible fall is produced, facing inward when the water in the ocean is highest, and outward

when the water in the harbor is highest. Hydraulic currents of this type are important features at the inlets connecting the ocean with lagoons behind offshore bars along much of the Atlantic coast.

Hydraulic currents greatly complicate the true tidal movements of coastal waters. An idea of their importance may be gained from an inspection of the review of tidal currents for different parts of the world given by Harris in his "Manual of Tides," where many of the associated hydraulic currents are mentioned[78]. According to Parsons the tidal currents in New York Harbor vary greatly in character, some being almost wholly oscillatory, others almost wholly hydraulic, and the remainder combining both elements in varying proportions[79]. Many, if not most, of the tidal currents observed along a coast are compound currents, consisting in part of true oscillatory movements of the water and in part of hydraulic movements. This is doubtless true of many of the tidal currents whose velocities are noted on a preceding page.

Seiche Currents. — The phenomena of seiches have already been described. It is evident that the rising and falling of water due to seiches in a lake, or in a bay of the ocean, must produce currents. As a rule these currents are so feeble in the main water body as to be scarcely perceptible; but if the waters are compressed into a narrower or shallower space, they may acquire an appreciable velocity. If the waters temporarily raised or lowered at one end of the basin are connected by a narrow strait with another water body, hydraulic currents of considerable force may be produced in the strait. The remarkable currents in the Strait of Euripus[80] appear to be largely of this origin. According to tradition Aristotle plunged into these turbulent waters in despair because he could not solve the mystery of their movements. The behavior of the water in the Strait is enough to justify the tradition, for the seiche currents are combined with tidal currents in such manner as to give nearly normal tidal movements for several days, followed by another period in which the waters ebb and flow twelve or fourteen times a day[81]. "The currents . . . are so violent that mills are kept in operation by them"[82]. Seiche currents must frequently modify tidal and other currents to an extent not yet determined; but it does not seem probable that they are often so strongly

developed as naterially to affect shoreline processes. Even in the Strait of Euripus, Cold[83] was unable to find any effect of the seiche currents upon the shores.

Wind Currents. — When wind blows over water it tends to drag the surface particles of the water along with it. Thus the water surface acquires a motion in the direction of the wind, although the velocity of the water never equals that of the wind. Because of the viscosity of water this motion is gradually communicated to the deeper layers but with rapidly diminishing intensity. It has been demonstrated that in course of time continuous wind action upon an unconfined ocean would set the entire body of the ocean in motion[84]. The surface currents produced by wind are often spoken of as wind drift, or drift currents; but since the term "drift" is also applied to currents of almost any origin which happen to flow parallel to the coast, as well as to shore detritus which is being moved by such currents, and since the terms "drift" and "drifting" are used in a restricted sense in this volume, it will be better for sake of clearness to employ the term "wind currents" when referring to the currents now under discussion. This term is not wholly satisfactory, as it suggests rather too strongly the air currents which are the cause of the water currents here considered; but since air currents cannot properly be called "wind currents," and since the term wind currents is analogous to the terms wave currents, tidal currents, and seiche currents already used, we may continue to speak of wind currents until a better term is suggested.

The velocity of wind currents will depend upon the strength of the wind, the length of time it has been blowing, and the size and shape of the water body. In the open ocean the surface waters under the trade winds ordinarily have a velocity of from 15 to 25 miles per day. Along the shore a velocity of three or four miles an hour during a strong wind is not unknown. Harrington[85] reports wind currents on the Great Lakes moving from 2 to 3 miles per hour, and Taylor[86] observed a current on the east shore of Lake Michigan which moved northward under the influence of "a strong sou'wester" with an estimated velocity of 4 miles. Currents of such velocities moving over a shallow bottom parallel to the shore are doubtless effective in the longshore transportation of débris, helping to remove eroded

material from the bases of cliffs and river-brought sediment from opposite stream mouths, as well as determining the character of the shores where their loads are deposited.

Hydraulic Currents Due to Winds. — Wind currents are extremely effective in causing hydraulic currents. If a wind current impinges directly upon a coast, the water is piled up above its natural level. In shallow water bodies, or on a shelving shore, the rise in level may be very marked, but is slight on steep coasts with deep water close in shore. The heaped up waters must escape to one side or along the bottom. If the latter mode of escape prevails, the seaward moving waters resemble the undertow, and may assist in the removal of fine débris to deep water. Such currents are sometimes spoken of as "counter currents in depth." Southeasterly winds in summer drive the surface waters of the Gulf of California northward toward the head of the Gulf, whence there is no opportunity for escape on the surface. At a depth of 50 meters the water is found to be moving southward[87]. Hunt observed a strong bottom current flowing out of Torquay Harbor when a gale drove the surface water inward[88].

The reverse of this circulation occurs when winds blow offshore, driving the surface waters out to sea. Bottom currents then move in toward the land to replace the water driven away by the wind. On the coast of Europe bathers are familiar with the fact that the water is warmer when the winds blow toward the land and colder when they blow in the opposite direction. This is because onshore winds pile the warm surface waters up against the coast, and the colder bottom water escapes seaward; while offshore winds blow the warm water away from the coast, and colder bottom water moves in to take its place. The northeast trade winds continually blow the surface water away from the northwest coast of Africa, with the result that abnormally cold water is always found near that shore, having moved in from the offshore depths[89]. Similar effects are produced in winter on the northeast coast of North America by the prevailing westerlies[90], and on other coasts which have prevailing offshore winds. Bottom currents of the type here described are probably comparatively feeble as a rule; but under favorable conditions, as in channels between shallows or in a shallow bay where the water blown in can only escape as a thin bottom layer, they may have velocities sufficient to move fairly coarse débris,

especially if the bottom is agitated by waves which keep the débris in suspension. Such débris would then migrate landward with an offshore wind, and seaward when the wind is onshore.

Important surface currents result when waters heaped up by the wind against a coast escape to either side along the shore, or through some strait into an adjacent water body. Water driven westward across the Atlantic by the trade winds piles up against the western shores of the Caribbean Sea, raising the level of the Sea above that of the Gulf of Mexico. There results an hydraulic current through the Strait of Yucatan into the Gulf which is one of the strongest of known ocean currents, having a velocity of 60 to 120 miles a day. The Gulf of Mexico, in turn, is higher than the Atlantic Ocean, and hence an hydraulic current passes through the Florida Strait into the ocean with a velocity of 70 to 100 miles a day. This is the beginning of the Gulf Stream proper, which off Cape Florida is only 15 miles from the shore and affects the bottom to a depth of nearly 3000 feet. " By calculation it has been shown that a current of the velocity of the Gulf Stream requires a difference of elevation of at least 0.7 feet of the Gulf over the Atlantic, which difference agrees very nearly with that found by direct leveling across the Florida Peninsula "[91]. A line of levels run between Cedar Keys on the Gulf coast and St. Augustine on the Atlantic indicates that the difference in level is probably at least 0.8 feet[92]. A current of 11 miles per day flowing eastward through the strait between Cape Horn and the South Shetland Islands illustrates how direct wind impact and hydraulic forces may act in the same direction to give a compound current; for the winds of this region, which drive the water eastward through the strait, also pile water up against the coasts of Chile and the South Shetlands, whence the escape for the hydraulic current is also eastward through the strait[93].

Temporary Currents. — In addition to the more or less permanent wind currents referred to above, there are temporary currents of considerable local importance which result whenever strong winds blow several days in a given direction. In the shallow zone along a coast the entire mass of water may have so strong a " set " in the direction determined by the wind that one can readily note longshore transportation of bottom débris. If the wind

blows directly on shore, the temporary head may be sufficiently great to produce hydraulic currents of no mean importance. Thus during a furious gale on the 15th of January, 1818, the water in the Kattegat " rose $5\frac{3}{4}$ feet above the common waterstand "[94]. Northwest gales raise the level of the North Sea on the coast of Holland 4 or 5 feet above the normal tide heights, while an easterly wind raises the waters of the Black Sea against the coast of Bulgaria two feet above the ordinary level. Mitchell[95] estimated that a northeast storm blowing the shallow water of Long Island Sound westward toward Hell Gate in New York Harbor caused a rise of 6 feet at the latter point, and of 4 feet on the open coast. During the severe storm of November 21, 1900, when the wind attained a velocity of 80 miles per hour at Buffalo, the lake level was raised 8.4 feet[96]. On the Zuyder Zee, with heavy west winds the water is lowered 8 feet on the west coast, and raised correspondingly with east winds. Owing to a heavy gale from the northeast in December, 1904, the water in the southern part of the Baltic Sea was raised from 8 to 12 feet above its normal level[97]. In the Galveston storm of September 8, 1900, the Gulf waters rose 20 feet and were the principal agent of destruction in the city. During the storm of October 5, 1864, on the coast of India, the water was raised 24 feet at Calcutta[98]. Such inequalities of levels must give rise to hydraulic currents of greater or less importance depending upon the configuration of the shoreline. According to Harrington[99] hydraulic currents on the Great Lakes, resulting from the disturbance of water levels by storm winds, attain a velocity as high as 240 miles a day.

Seasonal Currents. — Intermediate between the permanent currents resulting from such winds as the trades, and the temporary currents due to local storm winds, are currents which prevail during one season or another because of seasonal variations in the winds. Seasonal variations of current direction must in turn result in seasonal variations in the direction of débris transportation, and hence in the size, form, and position of beaches. Thus the beach on the southwest point of Baker Island, in the Pacific Ocean near the equator, migrates from one side of the point to the other with the seasonal change in the winds[100]. Seasonal currents consist of true wind currents and of hydraulic currents resulting from the piling up of the wind-driven water against the continents. We may form some idea of the probable importance of these

currents if we know the seasonal inequalities of water level along different coasts. Harris has summarized a number of cases, showing seasonal inequalities which measure from a few inches to several feet[101]. It appears from this summary that the northerly winds of winter blowing across the Gulf of Mexico produce low water at Galveston in February, while the southerly summer winds produce high water in October, the difference in height due to this cause being 1.5 feet. In winter the northeast monsoons blow the water away from the coasts of the northern Indian Ocean, and the southwest monsoons of summer raise the level, the difference varying from 1.8 to 3.2 feet. The southwesterly winds which prevail at Panama during much of the year raise the water against the northern shores of the Gulf of Panama 2 feet higher than the level which exists when the northeasterly winds of February blow the water away from those shores. In general, it is noted that sealevel is highest at most tidal stations in summer or autumn, and lowest in winter or spring; and since winds tend to blow from the ocean toward the land in summer, and from the lands out to sea in winter, it appears that the piling up of the waters against the lands in summer must be the principal cause of high water at that time, rather than an expansion of the oceanic waters due to high summer temperatures, which at most could scarcely raise the ocean level 0.1 of a foot[102].

It is evident that such large seasonal inequalities of level as have been noted above must be accompanied by currents which vary in direction or intensity, or both, with changes of the seasons. In the Strait of Bab-el-Mandeb at the mouth of the Red Sea there is a southeasterly current during the summer, because the summer monsoons of the northern Indian Ocean blow the surface water out of the Gulf of Aden, making its level lower than that of the Red Sea. In winter the current in the strait flows northwest, because the winter monsoons raise the Gulf level, and because the Gulf waters are less saline and therefore less dense than those in the Red Sea. The currents in the Strait often have a velocity of 30 to 40 miles a day, and as high a figure as $2\frac{3}{4}$ knots per hour, or 66 sea-miles per day, has been recorded[103]. These seasonal variations of the surface currents interfere with the circulation due to differences of salinity which would otherwise cause a constantly inflowing current on the

surface and as constant an outflowing bottom current. As with other currents produced by the wind, the direction and intensity of the seasonal wind currents, and of their resulting hydraulic currents, depend on a number of factors, among which wind velocity, water depth and shore configuration are of highest importance. There is a summer current northward through Bering Strait because southerly summer winds raise the waters of the shallow northern part of Bering Sea above the level of the Arctic Ocean, and the form of the shores offers a northward escape through a narrow channel[104]. In this case the current is probably in part a true wind current, although largely hydraulic. It will generally be found that near a coast the direct wind currents do not move in the precise direction of the wind, but are deflected by the trend of the shore which introduces more or less of the hydraulic element into the currents.

Planetary Currents. — There exist in the principal ocean basins gigantic whirls or eddies which are commonly referred to collectively simply as "the ocean currents." The principal cause of these great movements of the oceanic waters is now known to be the planetary wind systems which blow over their surface, although earlier students assigned a more important place to differences in oceanic temperatures. It should be noted, however, that while the planetary winds constitute the prime cause, and we may therefore appropriately call the currents "planetary currents," many other factors must be recognized in any full explanation of their origin. Taking the North Atlantic circulation as an example, we find the southern side of the great whirl driven westward by the trade winds, and the northern side driven eastward by the prevailing westerlies. But account must also be taken of the deflective effect of obstructing land masses; of the constantly operating force arising from the earth's rotation which tends to deflect currents toward the right in the northern hemisphere; of the hydraulic action resulting when the waters driven westward are piled up against the Caribbean and Gulf coasts; of the large amount of rain and river water added to the ocean in the Gulf region; and, if we follow Ekman[105] and Carpenter[106], of the water drawn into the Gulf by reaction currents (see below), and of the sinking cold water near the poles and rising warm water in low latitudes. It is simpler to treat currents of such complex origin in connection with the

individual elements which combine to form them; but the planetary currents are sufficiently distinct and well known to require brief mention as such.

Planetary currents may have a fairly high velocity under favorable conditions, as has already been noted for that part of the North Atlantic circulation called the Gulf Stream. In Florida Strait this current may reach a velocity of 4 miles an hour, which is sufficient to move large stones; and the current in the Strait of Yucatan has an even greater velocity. Such velocities are exceptional, however, and the bottom waters of even these currents move more slowly. Furthermore, planetary currents are usually located in deeper water far from the coast, and can therefore have little effect upon the shoreline. The swift current of the Gulf Stream in Florida Strait is some miles off shore and is separated from the land by another and slower current moving in the opposite direction. As Krümmel[107] has pointed out, even where currents of this type do come in direct contact with the land they are almost always completely overpowered by tidal or other currents of much greater importance in shoreline processes.

A good account of the former exaggerated ideas regarding the geological work of ocean currents will be found in Rühl's review of the literature relating to the "Morphologischen Wirksamkeit der Meereströmungen"[108]. While Rühl does not discriminate sufficiently between the different types of currents found in the ocean, it is evident that many of the reports to which he refers deal with planetary currents. Pechuël-Loesche[109] gives an interesting discussion of the conditions which render planetary currents unimportant agents on shoreline development, but tends to underestimate the transporting power of currents, and apparently does not distinguish sufficiently between the less important planetary currents and the movements due to tides and other forces which often have a very high degree of importance. Those tempted to ascribe shore forms to currents represented on charts or described in coast survey publications will do well to remember that currents so reported are usually studied several miles from shore where the water is deep enough to be important for navigation; whereas the shallow waters near the shore, of the highest importance to students of shore forms, are usually very imperfectly examined, if at all. The

fact that a certain current is observed several miles off a coast is no indication whatever that the waters near the shore move in the same direction. In shallow water, it is true, a planetary current may reach and scour the bottom; and it has been stated that such action is taking place under the Gulf Stream between Florida and Cuba, and on Blake Plateau southeast of Georgia[110]. Indirectly, these currents aid wave erosion by helping to distribute the finer waste of the lands far over the ocean floor in water so deep that it cannot readily be returned to the shore zone[111].

Pressure Currents. — The weight of the atmosphere on the surface of the ocean is about 15 lbs. to the square inch, or about $8\frac{3}{4}$ tons per square yard. It is evident that if atmospheric movements remove part of this weight in one place and increase it in another, the sea surface must rise where the pressure is partially relieved and sink where it is increased. Lubbock has shown that as a rule a rise of 1 inch in the barometer causes a depression in the height of high water amounting to 7 inches at London, and 11 inches at Liverpool[112], while Bunt has found that a similar barometric rise produces a depression of 13.3 inches in the tides at Bristol[113]. Since these differences of level are usually distributed over broad areas, under the continuous application of pressures which alter but gradually, they probably do not often cause currents strong enough to be perceived. But if two water bodies connected by a strait are subjected to unequal pressure, currents may be produced in the strait which have a fairly high velocity. Ekman has shown that if the barometer fall 30 millimeters over the Baltic, the result would be the same as if the water in the Kattegat had risen 4 centimeters. This would be sufficient under certain conditions to reverse the surface stream normally flowing out through the connecting strait, and to give a distinct current into the Baltic[114]. In the Gulf of St. Lawrence a difference in atmospheric pressure is said to produce a flow of water from the area of higher to that of lower pressure, and to produce currents through the inlets connecting the Gulf with the ocean. High pressure over the Gulf of Mexico when there is low pressure over the ocean outside appreciably increases the velocity of the Gulf Stream[115]. It is difficult in these cases to make sure that the effects noted are wholly due to differences in pressure and are not affected to some extent by winds blowing from areas of high toward areas of low pressure.

Marked differences of pressure, so distributed over water bodies of the proper form and arrangement as to favor the production of pressure currents, do not appear to be sufficiently frequent or sufficiently lasting to make these currents of more than local and temporary importance.

Convection Currents. — The warming of sea water causes it to expand and become lighter, while cooling causes greater density and hence increased weight. Therefore, if one portion of the ocean is warmed or cooled more than another, convection currents might be produced which would endeavor to restore a condition of perfect equilibrium. The planetary currents, as already noted, have been seriously regarded by some as mainly the result of unequal heating of the ocean. There is little doubt that a slow exchange of polar and equatorial waters is favored by temperature differences, the cold polar waters sinking and creeping equatorward in depth, while the warmed equatorial waters flow poleward over the surface. That portion of this motion due to temperature conditions is, however, extremely slow. Marked differences of temperature at sealevel exist only between regions widely separated; and the resulting differences in ocean level are very small, since the greatest difference of specific gravity that can arise in the ocean from differences of temperature is about as 1:1.0043[116]. Hence the convection currents which arise must be very feeble. It should be noted, furthermore, that the heat which tends to make sea water lighter by expanding it, also causes evaporation and thereby tends to increase the water's density. The effects of increased temperature may often be more than counterbalanced by the effects of evaporation. It is doubtful whether, even in the case of a strait connecting two bodies of water, the currents arising from temperature differences alone are ever sufficiently strong to be of importance in shoreline processes.

Salinity Currents. — The specific gravities of fresh water and sea water are very different, the relation at 15° C. being as 1:1.027, and at O° C. as 1:1.0283, if the sea water contains $3\frac{1}{2}$ per cent of salt[117]. It follows that anything which locally dilutes the sea water, or which locally increases its salinity, will produce currents which may have a very high velocity. The most important causes of dilution or increase in salinity of the sea are rainfall, the outflow of river water, and evaporation. It is

evident that these processes must produce direct changes of level in addition to changes of specific gravity. Rainfall and the outflow of rivers raise the sealevel, while evaporation lowers it. Such differences of level must result in currents which will combine with the currents due to differences of specific gravity to form a single system of circulation, in which the higher, lighter water flows toward the lower and denser water on the surface, at the same time that the denser waters move along the bottom toward the region of water less dense. We will call the currents of this system, salinity currents.

Rainfall is not equally distributed over the surface of the ocean. The equatorial rain belt has an excess of precipitation, and the same is true of higher latitudes where rains, snows, and melting ice contribute a large amount of fresh water to the sea. The two intermediate zones, from near the equator to about 40° north and south latitude, are characterized by deficient rainfall. There must be a tendency, therefore, for surface currents to move from both low and high latitudes toward the intervening areas of small precipitation. Such a movement in the open ocean would be comparatively slow, and must be largely masked by other currents of greater importance.

Strongly marked differences in density are produced when ice melts in the sea, and the resulting currents should be well developed in such regions as around the ice barrier of the Antarctic continent. Pettersson[118] and Sandström[119] have made special studies of such currents, and have shown that the melting ice dilutes the surface water and causes an outward or seaward surface movement. The water below the ice is cooled, its density thereby increased, and it sinks to the bottom and flows outward as a bottom current. Between these two there must result an inward moving zone of water which has been neither cooled nor diluted. Pettersson goes so far as to attribute an important part of the main oceanic circulation to ice-melting, an extreme view not shared by most oceanographers. Barnes[120] has investigated the value of ice-melting currents in enabling navigators to locate icebergs from a distance; but it has not yet appeared that currents of this origin are of importance in shore processes.

Fresh water poured into the ocean by a large river raises the sealevel at that point and lowers the density of the ocean water.

Surface currents tend to move out in all directions, and the bottom, denser water to creep in toward the river mouth. On an open coast the surface currents may be strong in the immediate vicinity of the river's mouth but at greater distances the movements must be relatively feeble. Where rivers empty into a gulf or bay the level may be so much raised as to cause a very strong current at the outlet to the ocean. The unusual strength of the Gulf Stream may in part be due to the large amount of water brought into the Gulf of Mexico by the Mississippi and other rivers. Even if the fresh water does not actually raise the level of the Gulf, it must prevent a lowering of that level by evaporation, and thus cause a virtual rise relatively to the ocean outside[121]. The surface currents moving from the Arctic Ocean through Denmark and Davis Straits into the Atlantic are probably in part salinity currents. "The considerable precipitation, the influx from several large rivers, and especially the small evaporation, all go to maintaining a rather low density for Arctic waters as well as an increased, but of course very small, elevation of the surface. . . . Doubtless a considerable amount of water passes as an undercurrent from the Atlantic into the Arctic through the straits east and west of Iceland"[122]. In the Gulf of St. Lawrence the waters have a lower density and higher surface than in the Atlantic, and a surface current of 2 knots per hour through Cabot Strait is attributed by Harris, in part at least, to this fact[123], although Dawson thinks the influence of the St. Lawrence River water upon currents in the Gulf is apt to be exaggerated[124].

Salinity Currents at the Mouth of the Baltic Sea. — The enormous influx of river water into the Baltic Sea causes that water body to be almost fresh at its northern end, and to have a low density throughout; while its surface is generally believed to be higher than the mean level of the sea outside. On this basis we should expect a surface current passing outward through the straits at the mouth of the Baltic, and an undercurrent of heavier, salt water flowing into the Baltic along the bottom. Such a circulation exists, and the velocity of the outflowing surface stream is usually given as 1 to 2 knots per hour in the Kattegat, but may be double this along the Norwegian coast of the Skagerack. It is strongest in spring and early summer, when the influx of fresh water into the Baltic

is at its maximum[125]. As Otto[126] has pointed out, unless prevented by other currents of greater power, such a circulation would result in the shore débris's being controlled by the outward flowing surface current, while the bottom débris in deeper water would be swept in the opposite direction by the inflowing bottom current. Ekman[127] attributes the deep channel in the Skagerack and Kattegat to the scouring action of the bottom current, which prevented deposition along its course of the sediment now covering the bottom of the North Sea; but he thinks, apparently without sufficient reason, that this was done when the land was higher and melting ice supplied larger volumes of outflowing waters. Perhaps a more probable interpretation is that the channel represents a normal river valley, submerged, and since kept open by current action.

Pettersson[128] has questioned the existence of a higher surface level in the Baltic on the ground that accurate measurements show the water level at Ystad and Landsort on the Baltic coast to be .024 and .023 meters respectively *below* the mean annual level at Varberg on the shore of the Kattegat; while Bjerknes and Sandström[129] contend that the difference in density between the Baltic water and that outside is not sufficient to account for the existing currents in the Belts and Kattegat. It should be noted, however, that the currents behave in a manner normal for salinity currents, and they are generally interpreted as such. The Black Sea receives every year 152 cubic kilometers (about 36 cubic miles) more fresh water than escapes by evaporation. Strong salinity currents therefore exist in the Bosphorus, the outflowing fresher surface stream at Constantinople having a velocity of 123 centimeters per second, or over 2 knots per hour. At a depth of 25 meters the heavier salt water is flowing inward with a velocity of 73 centimeters per second, the velocity decreasing slowly with increase in depth[130].

Salinity Currents at the Strait of Gibraltar. — Evaporation is an effective agent in producing salinity currents but in this case the surface current must of course flow inward toward the region of evaporation, where the water is increasing in density and the surface is being lowered; while the heavier salt water will flow outward at a lower level. A striking example of such circulation is found in the Strait of Gibraltar. The annual evaporation from the surface of the Mediterranean

amounts to a layer of water at least 3 meters deep according to Fischer[131], and greatly exceeds the influx of fresh water, with the result that the waters in the sea become denser and the surface lower than is the case in the Atlantic Ocean. The higher and lighter waters of the Atlantic flow into the Mediterranean as a surface stream of marked strength, while deep-water observations prove that a strong current of more saline water moves outward on the bottom. The great velocity of these currents is a matter of considerable interest. Maury[132] quotes the following from the abstract log of Lieutenant W. G. Temple for March 8, 1855, relating to the inflowing surface current: "Weather fine; made $1\frac{1}{4}$ pt. leeway. At noon, stood in to Almiria Bay, and anchored off the village of Roguetas. Found a great number of vessels waiting for a chance to get to the westward, and learned from them that at least a thousand sail are weatherbound between this and Gibraltar. Some of them have been so for six weeks, and have even got as far as Malaga, only to be swept back by the current. Indeed, no vessel had been able to get out into the Atlantic for three months past." It would seem from this that the surface salinity current, reinforced no doubt by an hydraulic current due to heaping up of water in the Gulf of Cadiz under westerly winds[133], and perhaps also to some extent by a direct wind current, had a velocity sufficiently great to prevent sailing vessels from passing westward to the Atlantic for months at a time. Helland-Hansen[134] has shown that tidal currents also affect the movement of the waters in the strait, the direction of flow at a depth of 10 meters even being reversed from its usual inward course for a brief period on the day of his observations. The maximum velocity of the inflowing current at a depth of 10 meters was on that day, 118 centimeters per second, or 2.3 knots per hour. On another day the velocity of the inflowing current at a depth of 5 meters was 150 centimeters per second, or nearly 3 knots per hour. At a depth of 46 meters the inflowing current had a velocity of 1.8 knots, and at a depth of 91 meters a velocity of 2 knots. The depth for the next series of observations was 183 meters (100 fathoms) and both here and below the current was continuously flowing out into the Atlantic. On the surface the current nearly always flows inward with a velocity of about 3 knots per hour[135].

The strength of the outflowing bottom current is more re-

markable than that of the inflowing surface current. With 274 meters (150 fathoms) of wire out the exact depth could not be learned because the wire was so strongly bowed by the force of the current. A maximum velocity of 227 centimeters per second, or 4.4 knots per hour, was recorded. When sent down with 366 meters of wire the apparatus was wrecked, apparently by being bumped against stones on the bottom[136]. Sir James Anderson has stated that the velocity of this outflow is so great at the bottom that at a depth of 500 fathoms the wire of the Falmouth cable near Gibraltar was ground like the edge of a razor, so that it had to be abandoned and a new one laid well inshore[137]. Captain Nares reports that he could get no specimen of the bottom, probably because of a "perfect swirl at that depth"[138]. Such currents must be very effective, not only in scouring the bottom at great depths, but also in transporting to a final resting place in very deep water any débris which may be delivered to it by the agitated surface waters.

Salinity Currents at the Strait of Bab-el-Mandeb. — Important salinity currents due to evaporation occur in the Strait of Bab-el-Mandeb at the mouth of the Red Sea. This sea is located in one of the dryest regions of the world, and possesses the highest mean annual salinity of any body of water in communication with the open ocean[139]. As a consequence there is an inflow of lighter water on the surface of the strait and an outflow of heavy salt water on the bottom, except when this circulation is interfered with by hydraulic currents caused by the monsoon winds. The velocity of the inflowing current is variously stated as from 30 to 65 knots per day, or a maximum of about $2\frac{3}{4}$ knots per hour[140]. The outflowing bottom current varies from 1 to 3 knots per hour. Even the lowest velocity mentioned for either stream is sufficient to move fine gravel; and it cannot be doubted that currents of this type play an important rôle along the shores of straits and the narrow parts of adjacent seas, even though the swiftest current is never found in the immediate vicinity of the shoreline.

River Currents. — Rivers entering the sea have their currents checked before they have advanced far into the quieter water, and in place of a narrow stream of fresh water moving forward under the impetus of the river's original velocity, there are developed slower hydraulic currents due to the piling up of the

waters, salinity currents due to differences in specific gravity, and reaction and eddy currents generated by the dynamic force of the original stream. For a short distance, however, one may recognize the true river current, the extent of its penetration as such into the sea depending on the volume and velocity of the river, the form of the shore and sea-bottom, and other factors. Dall[141] has attributed the clockwise circulation of the Bering Sea in part to river currents which enter the eastern side of the sea with a southwestward trend.

As the river current is checked by contact with the quieter waters of the sea it must of course deposit the débris it is transporting, the coarsest first, the finer as the current grows more and more sluggish. If the river is heavily laden with sediment, and the water body into which it empties is not greatly agitated by other types of currents, much of the débris will remain where first dropped to form a delta. Most rapid deposition occurs beneath and along the immediate margins of the river current, with the result that the current is ultimately confined between walls of its own deposits and prevented from coming in contact with the adjacent waters until it has passed beyond the limits of the embankments. Thus, the river current is carried farther and farther out into the oceanic waters between the two sides of an elongating delta lobe, as in the case of the Mississippi delta, some lobes of which have advanced into the Gulf of Mexico at the rate of from one hundred to several hundred feet a year. On the other hand, if a strong current of any type sweeps along the coast opposite the mouth of a river, the river current may be deflected so as to merge with the longshore current. As the river current gradually loses its identity the sediment is carried on by the higher velocity of the more powerful longshore current. Under these conditions no delta will form. Assuming that a river brings down a significant amount of sediment, its ability to form a delta does not depend upon its entering a tideless sea, as is usually stated, but rather upon its entering a comparatively currentless sea. The Indus builds its delta in a sea having a tidal range of 10 feet, while the Ganges delta has formed where the range is 16 feet. The theory that deltas are restricted to tideless seas is fallacious. If the river current is stronger than other currents in the sea at the point of its embouchure, and in consequence is carrying débris which those other currents can-

not transport, a delta will form, whatever may be the tidal range. If the strength of any other type of current exceed that of the river current at the point of embouchure, no delta will form.

It often happens that the river current is strongest at the immediate point of embouchure to begin with, although a more powerful current sweeps along the coast some distance out in the sea. This is especially apt to be the case where a river enters the sea at the head of a reentrant angle or bay. Delta formation will then proceed until the river current has been carried seaward, by the advancing delta lobe, to the point where it conflicts with and is overcome by the longshore current. Such seems to have been the history of the Nile delta; for although the river current brings out to sea 36,600,000 cubic meters of silt annually, this vast tribute of sediment does not add to the seaward extent of the delta, because " a powerful marine current sweeps past the coast and carries the sediment eastward beyond the most easterly mouth of the river "[142]. Much of the sediment now brought down by the Amazon is carried seaward with the aid of strong tidal currents, then caught up by the Northern branch of the South Equatorial current, which transports part of it over 300 miles to deposit it along the coast of Guiana[143]. The forms of deltas will evidently depend not only upon the original form of the shoreline, the nature and quantity of the débris brought down by the river, and the manner in which the river shifts its position upon the delta; but also upon the extent of wave action, and the direction and strength of coastal currents of different types as compared with the strength and direction of the river current.

Reaction Currents. — F. L. Ekman[144] has shown that since a river flowing into the sea sets adjacent water particles moving forward in the same direction, thereby increasing the volume of the current more rapidly than its velocity is decreased, there must be an influx of seawater toward the mouth of the river to make good the resulting deficiency. " Every river or brook which falls into the sea gives rise to an undercurrent directed toward its embouchure. These undercurrents are so distinct and the causes that produce them so active, that in calm weather their presence may be easily observed at the mouth of the most insignificant rivulet that falls out over the surface of the sea." To such currents Ekman gives the name " reaction streams."

Cornish has called them " induction currents "[145]. Investigations of the outlet of the Göta-Elf into the Kattegat showed that a reaction current flowed well into the bed of the river as a distinct bottom current of salt water. A sunken object was moved up the river channel by this current, in direct opposition to the surface flow[146]. It was shown that this current could not be explained as a mere salinity current due to differences of specific gravity between the fresh and salt water. Ekman even goes further, and regards the bottom currents at the outlet of the Baltic Sea and in the Strait of Gibraltar as in large part reaction currents. Cronander[147], on the other hand, would seem to doubt the existence of true reaction currents, even at the outlet of the Göta-Elf where Ekman made his principal study. While there are probably reaction currents developed both at the mouth of the Baltic and at the inlet to the Mediterranean, Ekman seems to push his theory too far and to lose sight of the facts that salinity currents of large volume must exist under the conditions obtaining at such straits as those in question, and that any reaction currents found there are secondary phenomena of less importance than the currents which give rise to them. Reaction currents have been further studied by V. W. Ekman, the son of the investigator quoted above, and some of his conclusions are embodied in a valuable paper[148] published in 1899. According to his studies, reaction currents are not always well developed at the mouths of rivers, and may even fail entirely[149]. On the other hand, Buchanan[150] goes so far as to explain the submarine gorge opposite the mouth of the Congo as due to reaction currents, which prevented sedimentation in the seaward prolongation of the river's course while the continental shelf on either side was being built up.

There can be little doubt that reaction currents must have some effect upon the transportation of débris in the vicinity of river mouths, and possibly in other localities. But while bottom débris has been observed in motion under the influence of these currents, our knowledge of their geological work and its relative importance is very slight.

Eddy Currents. — Closely related to the reaction currents described above are the eddy currents, which also result from the dynamic force exerted by the moving waters of currents of other types. In the typical reaction current the water moves

in under the original current which produced it. Eddy currents (called "draught currents" by Bache), on the other hand, are surface whirls in which the water next the original current moves forward beside it, the opposite side of the whirl flowing in the reverse direction. Thus the clockwise planetary whirls of the northern oceans give rise to counter-clockwise eddies on their outer sides. The surface manifestations of these whirls are so well known that it seems desirable, notwithstanding their close affinity with the reaction currents, to treat them separately under the name of eddy currents.

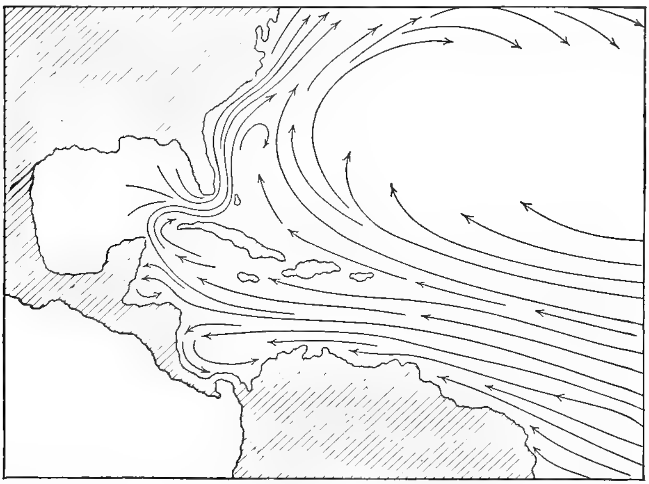

FIG. 21. — Eddy currents in the Gulf of Honduras and Mosquito Gulf.

The salinity current entering the Mediterranean Sea moves eastward along the northern coast of Africa, aided by the prevailing westerly winds. In the gulf off the coast of Tripoli it causes a well marked eddy current. The Equatorial Current flowing through the Caribbean Sea produces one eddy current in the Gulf of Honduras and another in the larger embayment of Mosquito Gulf north of Panama (Fig. 21). Tidal currents entering New York Harbor cause an eddy current just inside of Sandy Hook which must affect the development of that spit[151]. Gulliver has shown that eddy currents developed by tidal currents in estuaries may help to determine the detailed form of the shoreline[152], and Abbe has even attributed the formation of the

great Carolina capes to eddy currents generated by the Gulf Stream[153]. A great deal of importance has been attributed the Florida counter-current in determining the shore forms to along the eastern and southern coasts of that peninsula[154]. While there may be some question as to the origin of this current, and some even doubt whether its existence has been fully established, Perkins[155] is of the opinion that in so far as it is a reality it is probably an eddy current generated by the Gulf Stream.

Deflection of Currents. — All of the currents above described are subject to the deflective effect of the earth's rotation. Those in the northern hemisphere are deflected to the right, those in the southern hemisphere to the left. The deflection is unrecognizable in short, temporary currents, such as those arising from wave action; but is prominently shown by large continuously moving currents, like those of the planetary circulation, and may even be observed in the smaller salinity currents and other similarly restricted circulations.

COMPLEXITIES OF CURRENT ACTION

The preceding discussion of the several types of currents encountered in the sea is sufficient to show that the subject is by no means a simple one. We have endeavored to analyze the origin and nature of each type separately, and to gain some idea of its probable relative importance. But we fully realize that in nature one seldom encounters one of these currents operating alone. In almost every case the ocean water moves in a given direction because of the combined influence of several forces. At the Strait of Gibraltar the inward surface flow may at a given moment represent the combined effect of salinity, wind, hydraulic, tidal, pressure, and reaction currents, all moving in the same direction. Off Storeggen movements of the water toward the northeast were found to result from the combined action of planetary and tidal currents[156]. Along the south coast of Alaska a prominent planetary or eddy current and the local tidal currents are so far affected by wind currents that it has been asserted that "the currents along this part of the coast are controlled entirely by the winds"[157]. The currents in the Strait of Bab-el-Mandeb are variable in character because salinity, wind, and hydraulic

currents combine in varying proportions at different times of the year; and they are further " confused through the irregular tidal influence felt there "[158]. Tidal currents on the south coast of Cantyre are uncertain and imperfectly understood, being much affected by wind currents[159]. Ekman has described the complex nature of the Gulf Stream[160]; Buchan and Ekman have both discussed at length the combined effects of salinity and temperature on oceanic circulation[161]; and Parsons has described the combination of tidal and hydraulic currents in New York Harbor, and mentioned the difficulties arising from the interfering action of salinity, wind, and eddy currents[162]. Wind, pressure, and hydraulic currents may combine to reverse the normal outflowing salinity current at the mouth of the Baltic[163], while the similar current out of the Black Sea is reversed during strong south winds[164]. Cronander even goes so far as to reject the commonly accepted theory of salinity currents at the mouth of the Baltic, and regards both surface and bottom currents as due to the wind[165]. The continuous outflowing current just inside the northern end of Sandy Hook is part of the time a true tidal ebb current, and part of the time an eddy current developed by the flood tide[166]. Otto has shown the difficulty of analyzing the movement of shore and bottom débris by currents along the south shore of the Baltic[167].

Further complication arises from the fact that along the same shore different types of currents may act with very different strengths, and the same current may have very different power in two adjacent areas. In the shallow water close to the beach, wind and wave currents are extremely effective, while tidal currents may be scarcely perceptible. A few yards out from the same shore, in water of moderate depth, a tidal current may sweep with irresistible force, while the wave current will be too feeble on the bottom to move coarse débris. Divers have found that while large surface waves will not interfere with their work on the bottom, a tidal current may sweep so strongly over the same spot that it becomes impossible to stand against it[168]. Let us imagine that in such a case the wind current and beach drift is toward the east, while the tidal current runs toward the west. At the shore one observer notes that throughout the year the wind and waves invariably cause the shingle to be moved visibly eastward. Another observer finds that the only

known source of supply for the rocks from which the shingle is derived lies to the east, and hence concludes that tidal currents transport the material westward. Both are right, for the tidal current carries the shingle westward so long as it remains in deep water; but as fast as part of the material is moved into shallow water, or is thrown upon the beach by unusually large storm waves, it comes under the influence of the eastward directed forces; and it continues to move in this direction so long as it is not washed back into deeper water where the westward moving tidal current prevails.

Conflicting Opinions Regarding Current Action. — A brief examination of the literature is sufficient to show that in cases similar to the one supposed above, one observer has frequently denied the validity of another's interpretation at the same time that he maintained the correctness of his own. The engineers and other authorities in Great Britain have of necessity paid much attention to the problems of coast erosion and transportation; and if one looks through some of the papers on this subject published in "Minutes of Proceedings of the Institution of Civil Engineers," he will be surprised at the wide differences of opinion there expressed by different experts, on the question as to what agent effects the longshore transportation of sand and shingle. Discussions on this point cover many pages and sometimes required the entire time of two or more meetings for their consideration. According to the views expressed, both in these discussions and before other learned societies, the transportation of shingle is due "chiefly, if not entirely, to the action of wind waves" (J. Scott Russell); "to the effects of the ocean-wave or ground-swell" (J. N. Douglas), since "waves possessed sufficient power to move shingle at considerable depths" (Joshua Wilson), or even "at very great depths" (E. Belcher); whereas "very little was ascribable to action of the tide" (G. B. Airy), for "the tide current does not affect the depths of more than 12 or 14 feet" (E. Belcher), and "the tidal streams had not sufficient velocity to exercise any mechanical power whatever" (R. A. C. Austen). On the other hand, we have the opinions that "shingle could scarcely be moved by the heaviest waves at greater depth than three fathoms" (J. M. Rendel); the formation of the great shingle deposit of the Dungeness "should be attributed, principally, to the counter-current of the tides" (G. Rennie); and "at

Cahore the driftage is solely due to the flow-tide currents" (G. H. Kinahan), while the movement of another shingle beach was due to "submarine currents which had the power of carrying pebbles along the shore at great depths" (Joseph Gibbs). As Hunt[169] has pointed out, although "the action of waves on sea-beaches and sea-bottoms has been much discussed during the last fifty years, . . . there is scarcely an important point connected with the subject that is accepted without dispute, whilst not only the opinions, but even the recorded observations of skilled observers are often, to all appearance, in hopeless conflict."

Not only the cause of shingle transportation, but also such questions as whether large or small débris travels farthest, and under what conditions waves build up or destroy shingle beaches, are in dispute. According to Coode[170] large pebbles travel farthest because they move more readily than small ones; Redman[171] agrees to the greater travelling power of the large material, as does also Reade[172], who rejects Coode's explanation, however, and suggests one of his own. On the other hand Prestwich[173], Palmer[174], Airy[175], Spratt[176], and Geikie[177] hold that the smaller pebbles are those which travel farthest.

The question as to whether the shingle travels east or west on the great Chesil Bank of the south coast of England has long been disputed, with eminent authorities on both sides. Whether the largest or smallest pebbles tend to accumulate at the top of the beach has likewise been vigorously debated. Coode[178], Matthews[179], and Shield[180] state that with offshore winds the waves build up shingle beaches, while with onshore winds the beaches are cut away; but Kinahan[181] is of the opinion that the reverse is the case. Palmer[182] concluded that when more than ten breakers arrived in a minute the beach was eroded, when less than ten, the beach was built up; but Coode[183] declares that so far as a rule can be established it is that any number of breakers greater than nine per minute causes the building up of the beach, while seven or less produces erosion.

Reasons for Conflicting Opinions. — The remarkable disagreement which has been illustrated above is not so surprising when one considers the complex origin of currents in the sea, and the enormous variability of wave and tidal action along a coast. There can be no doubt that in some localities tidal currents play a more important rôle in the longshore transportation of sand

and shingle than do wave currents, beach drifting, and related forces; and it is equally certain that in many other localities the currents associated with wave action are more important transporting agents than are those of tidal origin. In still other localities it may be difficult to determine which of these two types of currents exercise a predominant influence upon the shoreline, or whether some other current may not be more important than either. The present writer entertains no doubt that as a whole waves are far more important agents of long-shore movement of beach material than are tides or other forces.

It does not appear that the conclusions of the authorities quoted above were based on any adequate analysis of the complex forces operating along the shore. On the contrary, in a large number of the instances cited conclusions were based on isolated observations in a limited number of places, and while these observations were usually made with skill and accuracy, they were utterly inadequate as a basis for general conclusions concerning such difficult problems as those encountered at the shoreline. Erroneous ideas as to the strength of certain currents have crept into standard textbooks, as for example Reade's conclusions regarding the strength of tidal currents near Gibraltar based on observations which really related to salinity currents[184]. This is inevitable, in view of the limited knowledge of ocean currents which exists even to the present time. Again, the resemblance between certain currents of different origin is so close that special care must be taken properly to distinguish them. Thus, at the mouth of a river we may have a landward directed bottom current which may be a salinity current, a reaction current, or a floodtide current, or all of these combined. Mitchell[185] describes such a landward current at the mouth of the Hudson River, and regards it as a true flood-tide current which creeps in along the bottom because it is heavier than the brackish water in the river. Harris[186] refers to this same current as one of the " counter currents at the bottom of the channel " caused by " a fresh-water stream discharging into the ocean," and refers to Mitchell's work apparently under the impression that Mitchell regarded the movement as a reaction current. It seems to the present writer that the conditions in this water body are distinctly unfavorable for the development of either salinity or reaction currents of large volume and appreciable velocity, and

that Mitchell's work demonstrated the tidal origin of the principal movement. The fact that a certain current flows landward along the bottom of a river channel and consists of heavier, more saline water than is found above it, does not mean that such current is *caused* by either the dynamic force of the river current or the difference in specific gravity between salt and fresh water.

Another source of difficulty in interpreting current movements arises from the fact that the currents usually observed are not always the ones which do the most work. Thus, the prevailing winds may cause almost continuous but weak wind currents and wave currents in one direction, whereas the greatest storms may cause short-lived but remarkably vigorous wind and wave currents in the reverse direction. More material may be moved, and moved a greater distance, by the latter currents than by the more continuous weaker ones. Hence, the direction for the dominant transportation of beach material is contrary to the prevailing currents. It has happened that in such a case one observer erroneously concluded that wave and wind currents had nothing to do with the distribution of the beach material; while another assumed that the material must move with these prevailing currents, and accordingly developed erroneous theories regarding the laws of shingle transportation.

A further cause of misunderstanding is the long time which waves and currents have taken to produce certain effects observed along the coast. To the geologist, who is familiar with the slow operation of the forces of nature, it seems that waves, at least, work with comparative rapidity. But the ordinary observer, and even the skilled engineer, may find it difficult to attribute the vast accumulations of sand and shingle on our coasts to forces which seem to him almost impotent in comparison with the great work accomplished. Such is the view repeatedly expressed by Wheeler in his volume on "The Seacoast." In the opinion of this eminent engineer, "a careful consideration of all the circumstances that attach to beaches can only lead to the conclusion that the results which have been attained must be due to other and mightier forces than those now in existence." "It is certain that the enormous mass of sand, which now covers the littoral of the sea and the beds of estuaries, cannot have been deposited by existing agencies." "The enormous accumulation of shingle known as the Chesil Bank . . . must have been

accomplished under conditions very different to those which now exist"[187]. The geologist, on the other hand, recognizes in these extensive shore deposits the effects of ordinary forces of nature continued for a very long period of time. There is nothing in the deposits described by Wheeler to excite wonder, except their extent; and large deposits may be made by ordinary forces working a long time as well as by extraordinary forces working a short time. I have examined some of the largest beach accumulations on the English and other European coasts, as well as those on the Atlantic coast of the United States, and see no reason to doubt that they have been produced by the same waves and currents which are still at work upon them.

Conclusions. – In the preceding paragraphs I have endeavored to give the reader some idea of the serious difficulties which confront the student of shore processes. It must be confessed, however, that it is much easier to describe the complexities of currents, and to point out the mistakes which are frequently made in interpreting them, than it is to solve those complexities in a given case and present a discussion which is so conclusive as not to be open to criticism. Nevertheless, it was essential that we should enter upon our treatment of shoreline forms with a broad view of the problems connected with wave and current action, and with some appreciation of the variety of the forces which operate at the shore in different places and at different times. We are now prepared to consider the development of shorelines more intelligently, even if we are not prepared to assert with positiveness the precise part played by different currents in shaping each portion of any given shore.

The time will come when our present limited knowledge of both wave and current action will be enormously extended by means of improved mechanical appliances. The movements of débris upon the bottom at considerable depths during wave action, concerning which we can only theorize at present, will be actually observed by special electrical apparatus. Wave currents and currents of other types will be studied by observing the exact movements of débris under their control. Limited areas of the coastal waters will be exhaustively studied, every detail of the currents analyzed with care under varying conditions, and the movements of débris determined with far greater precision than is now possible. While shoreline problems will never be

simple, the researches of the future will yield a body of facts which will enable the geologist and engineer of some coming generation to predict shore changes and plan harbor and coast defenses with an assurance which will contradict the assertion of the present maritime engineer, that the forces operating at the shore are among the forces of nature, "which are subject to no calculation." In the meantime we may take some satisfaction from the fact, which will presently appear, that a great deal may be learned about current action by studying the forms of beaches, since these often provide a more reliable indication of the dominant currents in a given locality than do any direct observations feasible at the present time.

RÉSUMÉ

We have reviewed the essential characteristics of the more important types of currents, and gained some idea of their relative strength, and comparative importance in shore processes. It appears that a great variety of wave currents operate in a most complicated and irregular manner, sorting and transporting débris in shallow water and on the beach in different ways depending on differences in outline of shore, angle of offshore slope, angle of wave approach, size of waves, kind of waves, and other factors. Tidal currents are scarcely less complicated, although developed on a much larger scale, and therefore more easily studied. Seiche currents, wind currents, planetary currents, pressure currents, convection currents, salinity currents, river currents, reaction currents, eddy currents, and hydraulic currents have all been considered; and we have found that some of them have a small degree of local importance only, while others are of wide-spread occurrence, or have a volume and strength which make them of very great significance. These currents are deflected from their initial courses by the earth's rotation; they combine with each other or counteract each other in the most complicated ways; they are not infrequently wrongly identified, and their manner of working and relative importance are often matters of dispute. Some knowledge of their behavior is nevertheless essential to an understanding of shore forms, and we may in turn expect to gain further knowledge of the currents themselves when we study the forms they have helped to produce.

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CHAPTER IV

TERMINOLOGY AND CLASSIFICATION OF SHORES

Advance Summary. — Before undertaking a systematic discussion of shoreline development it is important to adopt a satisfactory terminology for the topographic elements of shores, and to agree upon a classification of shorelines which shall serve as a guide throughout this treatment. These are the two tasks attempted in the present chapter. With the aid of explanatory diagrams the terminology used in this volume is made clear. The discussion of shore terminology leads inevitably to a consideration of the terminology of peneplanes of marine and other origins, and space is given to an inquiry into the proper significance of the terms plains, planes, and peneplanes. The classification of shorelines is next essayed. After a brief review of previous methods of classification, a genetic scheme is adopted in which four primary types of shoreline are recognized. These are described, their chief subdivisions named, and, where circumstances make it advisable, discussed at some length. It is further pointed out that each class or sub-class of shorelines passes through its appropriate young, mature and old sequential stages of development.

Terminology of Shores. — The line where land and water meet has been called the shoreline, the strandline, the coast line, and the water line. The terms shore, beach, strand, and coast are also loosely used with varying significance by different writers. " Shore " is defined by Gulliver[1] as the water area immediately seaward from the shoreline; by modern legal authorities, as the space between low water and high water; and by Wheeler[2], as the land area immediately above high water. " Beach " is sometimes used to denote the zone between low water and high water, or to denote the débris found between those limits, while others regard it as extending some distance below low water. " Coast " may mean the narrow strip immediately landward from the shoreline[3], or it may imply a much broader zone extending some distance inland. Ratzel[4] discusses at some length the varying

significance attached to the word coast by different writers. Evidently there is a variety of usage in naming shore features, even among scientific workers. It is essential, therefore, that we adopt a terminology to be used throughout this discussion, and an effort will be made to secure the required precision with the least possible departure from common usages.

At the margin of the sea there are typically found three or four distinct zones, each of which is characterized by certain peculiar forms due to deposition or erosion. The zones, the erosion features, and the features due to deposition must each be clearly distinguished and receive appropriate names (Figs. 22 and 23). The most important of the four zones extends from low water mark to the base of the cliff, whether large or small, which

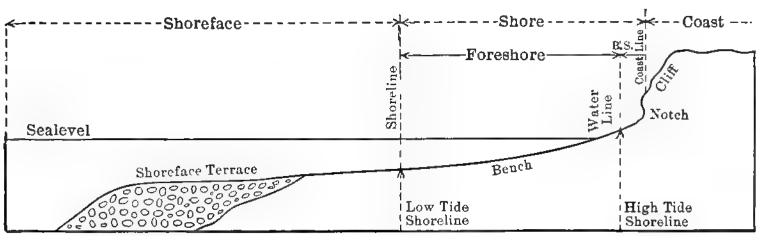


FIG. 22. – Elements of the shore zones during an early stage of development.

usually marks the landward limit of effective wave action. This is the zone over which the *water line*, the line of contact between land and sea, migrates; and it will here be called the *shore*. It is, indeed, the zone most commonly referred to when the word shore is employed in ordinary speech, and is likewise the zone defined as the shore in Roman law.

Landward from the shore is a much broader zone of indeterminate width, which will here be called the *coast*. While some may have more or less consciously included the shore when referring to a coast, it is also quite common to exclude it, by implication at least, as when one says that a coast terminates in a series of ragged cliffs. Indeed, the narrow shore zone is probably seldom thought of when a coast is referred to, and it will conduce to clearness if we restrict the terms shore and coast to the two independent zones. The line which forms the boundary between these zones is the *coast line*, and it marks the seaward limit of the permanently exposed coast. In a correspond-

ing manner the *low tide shoreline* marks the seaward limit of the intermittently exposed shore. The position of the water line at high tide marks the *high tide shoreline.* When the term " shoreline " alone is used in the text, low tide shoreline is to be understood.

Seaward from the low tide shoreline is a narrow zone permanently covered by water, over which the beach sands and gravels actively oscillate with changing wave conditions. Although of great importance to the student of shores, this zone has no suitable name. Gulliver[5] recognized this difficulty, and proposed to call the zone the " shore"; but his suggestion is hardly acceptable in view of the fact that " shore " is almost universally applied to some part of the land area inside the low water mark. Gulliver, furthermore, recognized only two zones, the coast and the shore. " Inshore " is sometimes used as opposed to " offshore," but is quite uniformly applied to a broader zone than that now under consideration; as, for example, in the expression " inshore fishing," which may refer to fishing carried on from one to three miles from the land. The term " shore face " as used by Barrell[6] in his discussion of deltas applies to much of the zone here in question, and after conference with that author I have decided to adopt his term, writing it as one word, *shoreface,* and redefining it as the zone between the low tide shoreline and the beginning of the more nearly horizontal surface included in the zone next defined. Extending from the outer margin of the rather steeply sloping shoreface to the edge of the continental shelf is a comparatively flat zone of variable width which will be called the offshore belt, or simply the *offshore* (Fig. 23). This is the zone commonly referred to in such expressions as " offshore sediments," and " offshore deposition."

The shore is sub-divided into two minor zones. One of these lies between the ordinary high and low water marks, and is daily traversed by the oscillating water line as the tides rise and fall. This zone is already well known as the *foreshore.* Back of it is the portion of the shore covered by water during exceptional storms only, which I propose to call the *backshore.*

The wave-erosion features associated with the coast, shore, shoreface, and offshore, are three in number. At the seaward edge of the coast is the wave cut *cliff,* which varies in magnitude

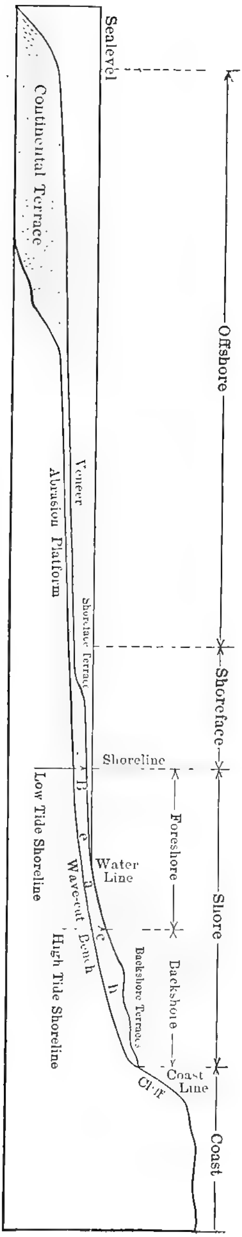


Fig. 23. — Elements of the shore zones in an advanced stage of development.

from an inconspicuous slope at the margin of a low coastal plain or in the side of a sand dune, to an escarpment hundreds of feet in height. In front of it, and occupying all of the shore zone and part or all of the shoreface is the wave cut *bench*, a sloping erosion plane inclined seaward. The bench may end abruptly at the top of a steeper slope representing part of the original surface of the sea-bottom (Fig. 22); or it may gradually decrease in slope until it merges imperceptibly into the more extensive, nearly horizontal plane produced by long continued wave erosion, which is commonly called the *abrasion platform* (Fig. 23).

There are three characteristic deposits which rest upon the wave cut bench and the slopes which lie to seaward of it. Most important of these is the deposit of material which is in more or less active transit, along shore or on-and-off shore, and which will be called the *beach*. Gilbert[7] defined the beach as "the *zone* occupied by the shore drift in transit." It seems to the present writer that the descriptions of beaches given by that careful and scholarly investigator of the topographic features of lake shores really deal with the deposits, and not with the zone in which those deposits occur; and what I have suggested is therefore a difference in phraseology rather than a real difference of interpretation. It would be unfortunate to have two names for the same zone, and none for the deposit which may or may not occur in that zone, as would be the

case if we accepted Gilbert's definition literally; for the zone over which material is in transit certainly includes the shore, and the transit of the material may either be so slow that some of it accumulates to form a deposit within the shore zone, or so rapid that bare rock is continuously exposed there (Fig. 22). Furthermore, the term "beach," as originally used, referred to the "shingle" or pebbles found on many of the English shores, and is employed in this sense in some parts of England to the present day.

Near the edge of the wave cut bench a portion of the beach is often fashioned into a terrace which is for a time progressively built out into the deeper water, only to be later modified or destroyed during heavy storms. This wave-built terrace may be called the *shoreface terrace* to distinguish it from a series of terraces caused by the action of storm waves on the upper part of the beach, and which we will call the *backshore terraces* (Fig. 23). As the abrasion platform is developed it may be covered with a thin deposit of material in slow transit, which constitutes the *veneer*. At the outer margin of the abrasion platform there accumulates an extensive deposit of the material which has been moved across the platform to a more permanent resting place in the deeper, quieter water beyond. This we will call the *continental terrace;* together with the abrasion platform it makes the *continental shelf.*

It will be shown in the following chapter that if a land mass stands still long enough, the waves will reduce it to an ultimate abrasion platform; and this is true, no matter how great may have been the original extent of the land. The final stage of shore development will witness the extinction of all of the above mentioned features, except only the abrasion platform and the continental terrace. Even the veneer may be removed and the bare rock surface of the platform exposed. Should an uplift raise the platform high above sealevel, stream erosion might dissect the new land area until only remnants of the former smooth erosion surface would be left. There are many dissected erosion surfaces in the world, some of which probably represent uplifted abrasion platforms. If the platform were reduced practically to a plane surface before uplift, the uplifted surface may be called a plane of marine denudation, or simply a *marine plane.* On the other hand, if wave erosion had not

yet succeeded in perfecting a smooth plane when uplift raised the platform above the reach of the waves to form a land area, the uplifted surface should be called a *marine peneplane.* These last two terms involve a slight revision of former usage, which cannot be appreciated without some consideration of the terminology of erosion planes in general. We may therefore turn our attention for a few moments to this broader subject.

Plains, Planes, and Peneplanes. — In an erosion cycle of any kind a land mass will in time be worn down to a smooth surface, providing the process of erosion is not interrupted. Long continued wave erosion reduces the land to a plane below sealevel; long continued stream erosion, to a plane at sealevel; and long continued wind erosion, to a surface at some elevation above the sea which will then be progressively lowered until sealevel is reached. Long continued glacial erosion may possibly produce a plane or a concave surface, either above or below sealevel; but the cycle of glacial erosion is not so well understood as are the other three cycles mentioned.

The work of any erosive force may be interrupted after the land has been worn down to a gently undulating surface of low relief, but before complete planation has been accomplished. We must recognize, therefore, not only the plane surfaces of ultimate erosion, but imperfect, " almost-plane " surfaces which characterize the penultimate stages of the several cycles. Theoretically, at least, there may be three, or possibly four, planes of erosion, with a corresponding number of almost-plane surfaces of uncompleted erosion, due to the action of rivers, waves, winds, and possibly glaciers.

Davis has given the name " peneplain " to the almost-plane surface of uncompleted fluvial denudation. The other almost-plane surfaces remain unnamed, except that such terms as " plain of marine denudation," " plain of marine abrasion," and " plain of aeolian erosion " have been applied to erosion surfaces, usually without regard to the question whether they were really plane surfaces, or only surfaces of moderate relief. An exception to the foregoing statement is perhaps to be found in Gulliver's valuable paper on " Shoreline Topography," where he seems to call the almost plane surface of marine erosion a " submarine platform," although he makes it identical with " plain of marine denudation," and would therefore seem to have no name for a

perfectly plane surface of marine denudation[8]. It does not seem advisable to apply the adjective " submarine " to a surface which is today far above sealevel, so some other name should be sought for up-lifted surfaces of marine erosion.

It is coming to be recognized that some of the so-called peneplains are more probably almost-plane surfaces of marine erosion, as was, indeed, the opinion of the earlier geologists. Barrell[9] has even questioned the subaërial origin of the New England uplands, supposedly a typical peneplain, and the one above which rises the mountain selected as the type " monadnock." Should his conclusions prove correct, and apply to the portion of the supposed peneplain in southern New Hampshire, then not only would the upland cease to merit the term " peneplain," but Mt. Monadnock would no longer be a " monadnock," as that term is generally defined. We would have to change the definition of monadnock, or invent a new name for the topographic feature of which it is an example.

A further difficulty arises from the fact that the word " plain " is used for two such very different conceptions as a plane of ultimate erosion, and a series of low-lying horizontal sediments which may be dissected into hills and valleys by stream erosion. A peneplain is not " almost a plain " of the second type, but is almost a plane surface in the mathematical sense of the term.

It appears, therefore, that we need names for the different types of erosion surfaces which theoretically may be produced both in the penultimate and the ultimate stages of erosion; that we also need a name for the actually existing upland surfaces of erosion which are so abundant in different parts of the world but of whose origin we are not yet certain; and, finally, that we need a clearer distinction in the names themselves, between " plains " of erosion, " plains " of deposition, and " peneplains." I believe it is possible to meet all these needs without departing unduly from the path of conservatism; and to this end the following usage will be adhered to in future pages: (1) The level erosion surface produced in the ultimate stage of any cycle will be called a *plane*. (2) The undulating erosion surface of moderate relief produced in the penultimate stage of any cycle will be called a *peneplane*. (3) A low-relief region of horizontal rocks will be called a *plain*. We may then

recognize planes of fluvial,* marine, aeolian, and glacial denudation; also fluvial peneplanes, marine peneplanes, aeolian peneplanes, and glacial peneplanes. A monadnock may be defined as an erosion remnant left standing above a peneplane of any origin, either because it is composed of more resistant rock or because it has been less exposed to the agents of erosion. Vogt[10] has already used the term "Monadnock-Berge" for isolated hills on the uplifted marine abrasion plane of northern Norway. It seems desirable to employ a special term for a surface of marine denudation which is still in its original position at or near wave base, with the marine forces still operating on it, and for this feature the name abrasion platform has already been used. An uplifted abrasion platform of large areal extent is a marine peneplane or a marine plane, according to the smoothness of the surface produced by wave erosion.

The advantages of this usage are obvious. It tends to simplify, not to complicate physiographic terminology. The origin of any of the level or almost level surfaces here discussed is at once apparent from the spelling. If in describing a coastal region one uses the otherwise non-commital term "marine plain," in the manner here suggested, the reader knows at once that a coastal plain of marine sediments is referred to; while "marine plane" indicates a wave-cut plane surface. When we remember that both of these forms have been called "marine plains," the advantage of the distinction in spelling is evident. One of our ablest geographers has applied to a wave-cut rock bench along the coast the term "coastal plain." Had he written "coastal plane" his meaning at least would have been clear, even though the term as a whole might still be criticized. A peneplain is not "almost a plain" of horizontal sediments, but is almost a plane surface in the mathematical sense of the term;

* "Subaërial" denudation was long used for "fluvial" denudation, because of the prevalent idea that there were only two important types of denudation, subaërial and submarine. Since the importance of aeolian denudation, which is also subaërial, has been recognized, it is desirable to distinguish the two types of subaërial denudation by the terms "fluvial" and "aeolian." It should be understood that the term fluvial is here used in its broadest sense, to include the action of rain water assisted by weathering and all other forces causing streams of water and waste from the largest to the smallest dimensions. Fluvial denudation is in reality pluvio-fluvial denudation.

therefore, " peneplane " more clearly expresses the true meaning of the term than does the older and commoner spelling. So standard a text as Dana's " Manual of Geology " employs the spelling " peneplane "[11], while J. W. Gregory[12] in his recent book on " The Nature and Origin of Fiords " makes use of the same form. Lawson has employed both spellings, the form "peneplane" occurring in his " Post-Pliocene Diastrophism of the Coast of Southern California "[13]. It may be, also, that the combination of "plane" with "pene-" will seem less objectionable to those who dislike hybrid terms, since " plane " is closer to the original Latin form than " plain."

Davis's introduction of " peneplain " into the vocabulary of physiography in 1889[14] was a valuable service to the science; for it led to a speedy and world-wide recognition of a conception which had previously been announced by Marvine[15] and extensively developed by Powell[16] and Dutton[17], but which did not become current until well named. It should be appreciated, however, that at this time the idea of subaërial denudation was supplanting, rather than supplementing, the idea of marine denudation. The general attitude was well expressed by de Lapparent[18] in the words: " La notion des pénéplaines est extrêment féconde, et ce n'est pas un de ses moindres mérites d'avoir porté le coup de grâce à la théorie des plaines de denudation marine, si fort en honneur de l'autre côté du détroit." Davis himself regarded extensive planes of marine denudation as " very improbable "[19], while planes of glacial or aeolian denudation were as yet scarcely considered. All planes of penultimate stages of erosion were called " peneplains," because it was believed that they were all formed essentially by subaërial agencies. No need was felt of names for almost-plane surfaces of marine and other types of erosion, since the existence of such planes was either considered improbable, or was not considered at all.

Later years have witnessed the publication of Passarge's studies of surfaces of aeolian denudation[20], and the probability of the existence of fairly extensive surfaces of marine denudation seems now to be recognized by Davis and others[21]. We are confronted with the fact that there are numerous all-most plane erosion surfaces in various parts of the world, the origin of which is in most cases doubtful, and in many cases will probably never

be known because nearly all of the upland has been destroyed by subsequent stream erosion. As Davis once remarked, "It must be remembered that the terms ' plain of marine denudation ' and ' peneplain ' are in nearly all cases hardly more than different names for the same thing. If the whole truth were known, it is probable that one or the other name might be appropriately applied in this or that case, but it is seldom that anyone has succeeded in convincing all his contemporaries that he could distinguish a plain of marine denudation from a peneplain, or vice versa "[22]. Present needs can better be met by applying the excellent term " peneplane " to all these surfaces, and qualifying the term by the word fluvial, marine, aeolian or glacial in case it can be shown that a given surface is of the origin indicated by the qualifier, rather than by inventing a new term for almost-plane erosion surfaces of doubtful origin. Peneplane is too valuable a term, and is too extensively used, to have it restricted to the very few (if any) erosion surfaces demonstrably of fluvial origin; and since the wise physiographer must avoid a name which commits him unwillingly to a certain theory of origin, it is best in the present case to extend the meaning of the term.

While it is true that I am advocating a broader significance for " peneplane " than is usually given to it, precedent for such usage is not altogether lacking. H. E. Gregory[23] is responsible for using " peneplain " as synonymous with " plain of denudation . . . carved out of other land forms either by the action of the forces that work on the land or by the waves of the sea." Davis himself has occasionally employed the term in the broader sense. Thus, in an account of the geographic development of Northern New Jersey, he discusses at length " whether the old Highland peneplain was the product of subaërial or submarine processes "[24]. He distinguishes between " subaërial base-level plain " and " submarine platform," but applies the name " peneplain " to the topographic feature itself while its origin remains in doubt. In a discussion of the arid cycle by the same author[25] " peneplains " are usually contrasted with true " plains " of aeolian erosion; but the frequent use of the expression " normal peneplain," and the application of the term " monadnock " to residuals on a plane of aeolian denudation, suggest that the author at least unconsciously recognized the possible existence of " peneplains " which were not formed by

" normal " (stream) erosion. Marine peneplanes are more definitely recognized, at least as a theoretical possibility requiring discussion, when in an essay on planes of marine and subaërial denudation the author says: " By whatever process the so-called ' plain of denudation ' was produced, an explanation that will account for a peneplain of moderate or slight relief is all that is necessary "[26].

Recognition of the fact that wave erosion is capable of producing a marine plane, or at least a marine peneplane, is essential to a full comprehension of some significant phases of shoreline activity. This matter will claim our attention in the next chapter.

Classification of Shorelines. — Various classifications of shorelines or coasts have been proposed, some of which are based on form rather than genesis, while others take account of the origin of shorelines but do not consider the stages of development reached since they originated. The first type of classification is wholly empirical and therefore not very significant; the second type is partly genetic, but not evolutionary, and is therefore less significant than it might be. Neither type permits one to arrange all shore forms in genetic series according to their relative advance in the cycle of shoreline evolution. Good examples of such classifications will be found in Suess' " Das Antlitz der Erde "[27], von Richthofen's " Führer für Forschungsreisende "[28], and Penck's " Morphologie der Erdoberfläche "[29]. Those desiring to study further the methods of classification here referred to will profit from a reading of Fischer's paper entitled "Zur entwickelungs-geschichte der Küsten "[30], and Hahn's paper on " Küsteneinteilung und Küstenentwickelung im verkehrsgeographischen Sinne "[31]. Applications of such methods in the description of specific coasts have been made by Haage[32] in his dissertation on " Die Deutsche Nordseeküste," by Meinhold[33] in an essay entitled " Die Küste der mittleren atlantischen Staaten Nordamerikas," and by Weidemüller[34] in his account of " Die Schwemmlandküsten der Vereinigten Staaten von Nordamerika." American students will be especially interested in the last two papers, since they relate to our own shores and at the same time furnish good examples of a type of physiographic description very commonly encountered in German writings. The details of coastal features are empirically described with painstaking care and at great length, instead of being represented by maps;

and while the origin of the features so described is then considered, each coast section is for the most part treated as a special isolated problem, without regard to its position in a series of sequential forms.

Numerical Methods. — Many attempts have been made to express the distinctions between different types of coasts in numerical terms. This method has been much in vogue among German students since the time of Ritter. The essential object of the method is to establish a comparison between coasts showing different degrees of indentation by the sea, and the comparison is usually expressed by a numerical relation rather than by absolute figures, such as the actual length of shoreline. Various relations have been utilized in this connection, such as the ratio existing between the length of the actual shoreline and the shortest possible shoreline which the area in question could have (Nagel); or the ratio of actual shoreline length to the length of an ideal contour connecting the outer points of the peninsulas, or the innermost points of the bayheads; or the ratio between the area of the main continental mass and the area of the outlying peninsulas and islands (Klöden). Ritter[35] and Berghaus[36] compared the area of the land with the length of the bordering shore, a method which was criticized by mathematicians on the ground that planes could not properly be compared with lines. Nagel[37] determined the size of a circle containing the same area as the land whose coast was under examination, and then compared the length of actual shoreline with the circumference of this circle.

Schwind[38] employs a comparison between the length of actual shoreline and the length of a selected isobath having a much simpler form, thus following the method of his teacher, Ratzel[39], who considers that the length of the shoreline should always be compared with some real contour-line. De Martonne[40] presents a number of valid criticisms against all these methods, and suggests that more significant results can be obtained by comparing lengths of shoreline with areas included between selected contours above and below sealevel.

These are but a few of the great number of schemes devised by different students in an attempt to discover a method which would not be open to the criticisms urged by each student against all the methods of his predecessors. We must restate today Hahn's conclusion[41] of a third of a century ago, that all numerical

methods of coast description are failures. All of them are essentially empirical, and hence of little or no significance to the student of shore forms. They tell little which a good map does not tell much better. Even when the numerical expression is combined with a discussion of the relation of the shoreline to geological structure, changes of level, the progress of shore accretion, and other phenomena affecting the coast, the result is a description only partially genetic, and one which fails to recognize the importance of shore processes in developing the shoreline in orderly, sequential stages.

The reader who would examine further the numerical schemes for coastal description will find a good historical review of the development of the method in Riessen's "Überblick und Kritik der Versuche Zahlenausdrücke für die grössere oder geringere Küstenentwickelung eines Landes oder Kontinentes zu finden"[42] while Schwind's essay on "Die Riasküsten und ihr Verhältnis zu den Fjordküsten unter besonderer Berücksichtigung der horizontalen Gliederung,"[43] contains a bibliography of the subject. Reference should also be made to the paper by de Martonne[44] already mentioned and to other papers by Reuschle[45], Schröter[46], and Hentzschel[47] where special applications of the method are considered. A good idea of the ingenious but highly artificial and complex mathematical methods of describing coastal embayments employed by Weule, Güttner, and others may be secured from Güttner's dissertation on "Geographische Homologien an den Küsten mit besonderer Berücksichtigung der Schwemmlandküsten"[48]. Descriptions of the southeastern coast of the United States based in part on numerical methods will be found in reports by Weule[49] and Weidemüller[50].

Genetic Classification of Shorelines. — The character of any shoreline must depend in the first instance upon the character of the land surface against which the sea comes to rest. If the surface is a partially submerged, irregularly dissected land area with numerous hills and valleys, the water will enter the branching valleys and spread around the hills, forming a very complicated shoreline. If the surface is a smooth, emerged sea-bottom, the shoreline is necessarily simple. It follows from this that the most significant classification of shorelines will be one which takes account of the nature of the movements of land or water which brought the water surface against the land at the present level.

It was upon such a genetic basis that Davis[51] divided shorelines into two primary classes, and that the more detailed discussion of Gulliver[52] was founded. There are, however, important shorelines which find no satisfactory place in the classifications of Davis and Gulliver, and for which provision must be made.

We will find it profitable to divide shorelines into four main classes: I, *Shorelines of submergence,* or those shorelines produced when the water surface comes to rest against a partially submerged land area; II, *Shorelines of emergence,* or those resulting when the water surface comes to rest against a partially emerged sea or lake floor; III, *Neutral Shorelines,* or those whose essential features do not depend on either the submergence of a former land surface or the emergence of a former subaqueous surface; IV, *Compound Shorelines,* or those whose essential features combine elements of at least two of the preceding classes.

Shorelines of submergence have been called "shorelines of depression"; but this implies a depression of the land, whereas the submergence may as well result from a rising of sea or lake level, or from the melting of tidewater glaciers permitting submergence of glacial troughs without any change in the relative level of either land or water. The term "shorelines of emergence" is likewise preferable to "shorelines of elevation," not only because emergence may result from the lowering of sea or lake surface, but for the added reason that "shorelines of elevation" and "elevated shorelines," two wholly distinct forms, are in danger of having their similar titles confused even though there is no possibility of confusing the forms themselves. The terms "sinking" and "rising" have been applied to coasts bordered by shorelines of submergence and shorelines of emergence; and in his admirable treatise on "Die Erklärende Beschreibung der Landformen" Davis[53] classifies the coasts, rather than the shorelines, into Senkungsküsten and Hebungsküsten. Such terms are open to the objection that they not only imply an actual change of level, and that it is a land movement which effects this change, but also that the movement is still going on; three implications which are probably not justified in the case of many coasts to which the terms are applied. We might employ the terms "positive shorelines" and "negative shorelines," thus indicating that the shorelines resulted from positive or negative movements of the water line, without indicating whether such movements

resulted from changes in the level of the land or the water. But these terms are open to two objections: they have been applied to land movements as well as to strand movements, and many are confused by the necessity of remembering that a positive land movement means a negative strand movement, whereas a positive water movement means a positive strand movement; furthermore, they imply that some vertical change in either land or water is necessary, whereas we have seen that submergence may occur with no change in the level of either. "Irregular shorelines" and "simple shorelines" are unsatisfactory, both because they are empirical terms which carry no suggestion of origin, and because all classes of shorelines are simple in mature or late mature stages of development. "Shorelines of submergence" and "shorelines of emergence" are explanatory terms; they are genetic rather than empirical; they do not carry any implication as to whether it is the land or the sea which moves, and do not even imply any vertical change of level in either; they are easily understood, and are not in danger of being confused with other terms applied to shoreline phenomena. For these reasons it seems desirable to use them instead of the other terms which have been mentioned.

I. Shorelines of submergence may be subdivided into two main types, according to the nature of the forms submerged. These are: (*a*) shorelines formed by the partial submergence of a land mass dissected by normal river valleys, which may be called *ria shorelines,* after the ria coast of northwestern Spain, which was produced by the drowning of normal river valleys along a mountainous coast; thus used, the term ria is not restricted to the narrow meaning assigned to it by von Richthofen, who first used it in a generic sense; but is employed in the broader sense in which it has been used by Gulliver and others: (*b*) shorelines produced by the partial submergence of a region of glacial troughs, known as *fjord shorelines,* like those of Norway and Alaska.

(*a*) *Ria Shorelines.* Since ria shorelines are produced by the partial submergence of normally dissected examples of the three main groups of constructional landforms (plains-plateaus, mountains and volcanoes), we may recognize as three subtypes: embayed plain or plateau shorelines, such as are found in the Chesapeake Bay region (Fig. 24); embayed mountain shorelines, of which the Maine coast and the coast of Brittany furnish good ex-

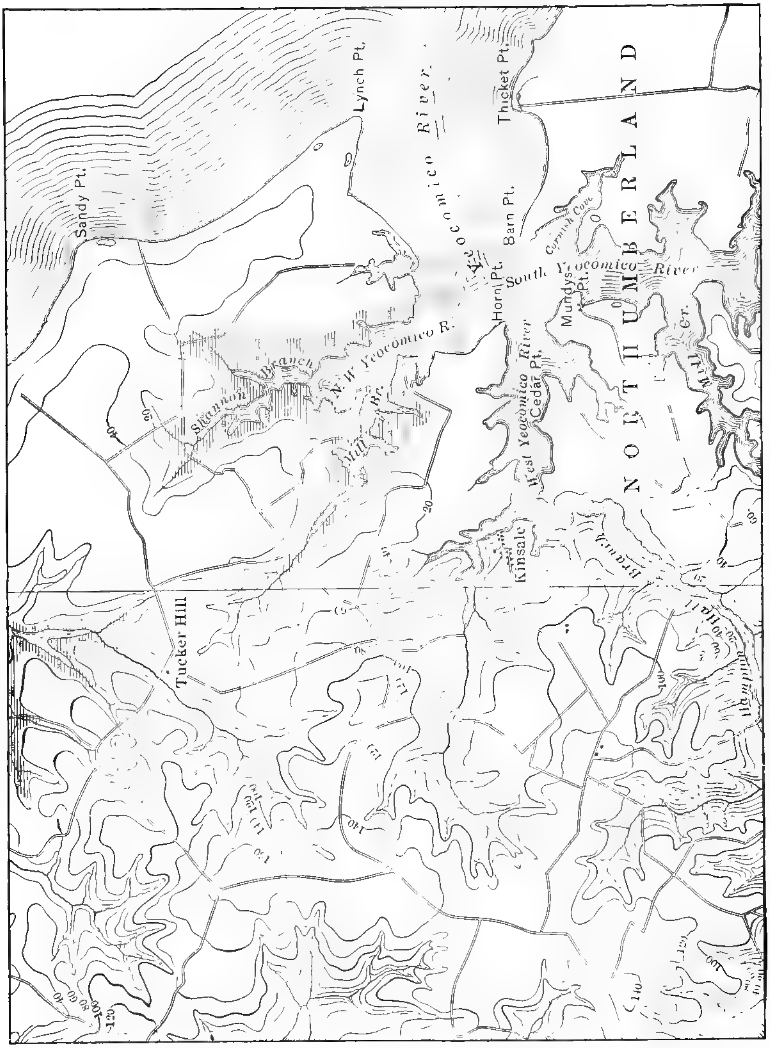
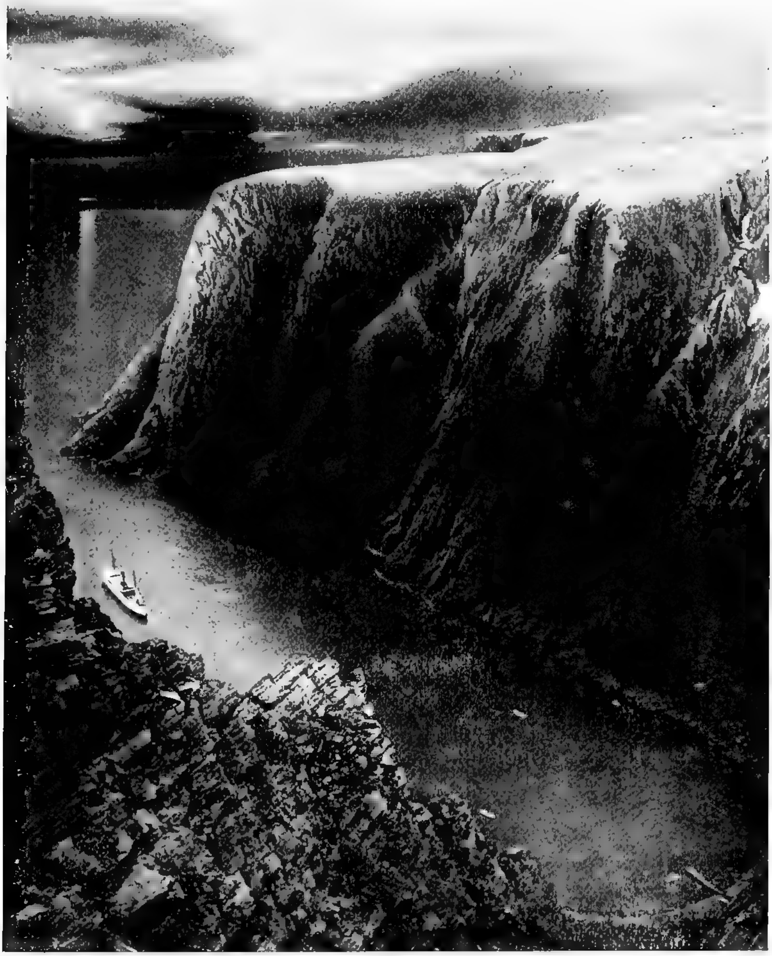


FIG. 24. — Embayed coastal plain of Chesapeake Bay region, showing example of ria shoreline.

PLATE XVI.



Hornviken, a small fjord near North Cape, Norway, showing typical over-steepened walls of a glacial trough.

amples; and embayed volcano shorelines, a number of which are found in the South Pacific, having been cited by Dana[54] in support of Darwin's theory of a subsidence of that ocean's floor. When desirable, the form of the shoreline may be more clearly indicated by specifying the particular type of plain, plateau, mountain, or volcano involved. Thus one may speak of the embayed coastal plain shoreline of Maryland, or the embayed folded mountain shoreline of the Adriatic coast of Dalmatia, thereby taking due account of the structure of the thing submerged, which is an important element in determining the character of any shoreline of submergence.

Care must be taken to avoid the ancient fallacy that the branching bays of ria coasts are due to wave and tidal erosion[55]. It is illustrative of the slow progress of the science of shorelines that Playfair, at the same time that he made his keen observations on the nature and origin of river valleys, should ascribe deep gulfs and salient promontories to differential wave erosion,[56] and that nearly a century later Fischer[57] and other students of shorelines should be found still advocating this view. Although it was recognized some time ago that as a rule tidal currents merely ebb and flow through submerged branching river valleys which they had no power to originate and which ordinarily they have but very moderate power to enlarge, and that wave erosion tends to obliterate the larger irregularities of a coast and not to make them, one still finds the tidal and wave origin of such drowned valleys as those of the Maine coast maintained in recent editions of a standard textbook on geology[58].

(*b*) *Fjord Shorelines.* — Perhaps no type of shoreline has given rise to so much discussion as has the fjord shoreline. We may note in the first place that geologists and geographers may be divided into two main groups whose ideas regarding the origin of fjords are mutually opposed. The first group may be designated as the "glacialists," because in their opinion all the phenomena peculiar to fjords may be explained as the result of extensive glacial over-deepening of pre-glacial river valleys near the sea. The second group, or "non-glacialists," reject the theory of ice erosion, and attempt to account for the phenomena of fjords in other ways.

According to the glacial theory, fjords are partially submerged glacial troughs. The troughs of glaciated mountains far from

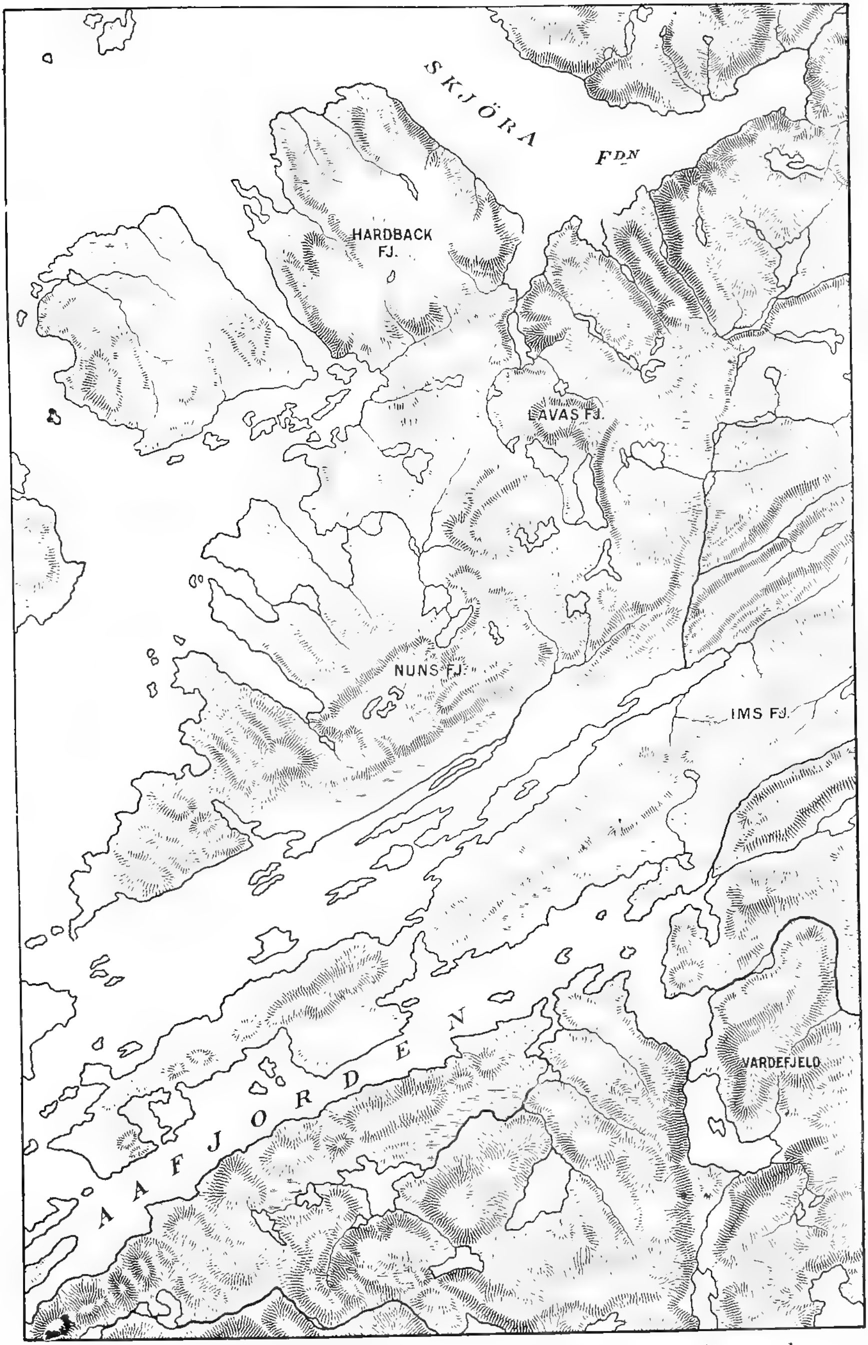


FIG. 25. — A typical section of the fjord coast of Norway, showing angular pattern attributed to fault-control.

PLATE XVII.

Photo by Underwood & Underwood.

Idde Fjord near Fredriksten, Norway, showing rectangular pattern characteristic of many fjord coasts.

the sea are similar to fjords, except that the former have not been drowned by marine waters. In both cases the troughs were formed by extensive glacial over-deepening of former river valleys. The pre-glacial valleys guided the glaciers which later came to occupy them, and by confining the ice streams to the narrow limits imposed by the valley walls, insured a maximum efficiency of glacial erosion. The glacial theory asks no questions as to what determined the courses of the pre-glacial valleys; but it is fully recognized that among other causes ancient fault lines must be considered, since a fault may give a crushed zone which is weaker than the unfractured rock, or may bring a belt of weak rock into such position that subsequent valleys will soon be excavated along it, parallel to the fault. This would satisfactorily account for the fact that many fjord shorelines have a more or less angular pattern. (Fig. 25 and Plate XVII).

Esmark[59] was the first to advocate the glacial origin of fjords, almost a century ago. The fjord valleys of New Zealand were ascribed largely to ice erosion by von Haast[60] in 1865, while Helland[61] a few years later, in discussing the fjords of Norway and Greenland, gave the best exposition of the glacial theory as applied to the interpretation of fjords which had appeared up to that time. Helland seems to have anticipated Shaler in recognizing the ability of glaciers to excavate their channels below sealevel, and to have given a fairly good account of the essential significance of hanging valleys some twenty years before Gannett's classic statement. The influence of rock fractures on the orientation of fjord valleys was recognized by Brögger[62], who did not fail, however, to attribute the actual excavation of the fjords to glacial erosion. In a similar manner Reusch[63] for the Norwegian fjords and Andrews[64] for those of New Zealand, make a clear distinction between the rôle of faulting in determining lines of weakness favorable to rapid stream and glacial erosion, and the rôle of glaciers in giving to the fjords their present form and depth.

In 1895 Shaler[65], in discussing changes of sealevel, accepted the glacial origin of fjords and stated that since glaciers may cut their channels below the surface of the sea, the flooding of a glacial trough may be accomplished as the ice melts, without any sinking of the land or rising of the water level. This same view, that fjords do not indicate past changes of level, was adopted

PLATE XVIII.

Photo by Underwood & Underwood.

The Naerö Fjord, Norway, a partially submerged glacial trough.

by Hubbard[66] in a brief review of the fjord problem which he published in 1901; by Daly[67] in his account of the Labrador fjords; and by Andrews[68] in discussing the fjords of New Zealand. It is further elaborated by Gilbert[69] in his report on glacial studies, forming the third volume of the Harriman Alaska Series, where the reader will find a discussion of the physics of glacial erosion below sealevel. Marshall[70] in his "Geography of New Zealand," and Tarr[71] in his report on the "Physiography and Glacial Geology of the Yakutat Bay Region, Alaska" are among other students of fjords who attribute their excavation to ice erosion.

Members of the non-glacialist group are by no means in agreement among themselves as to the origin of fjords. They agree on one thing only — that ice did not excavate these deeply submerged canyons. Some consider fjords the product of normal stream erosion followed by a partial submergence which permitted the valleys to be drowned. This was the view expressed by Dana[72], who first emphasized the restriction of fjords to high latitudes but did not suggest for them a glacial origin. Upham[73] definitely rejects the glacial explanation, and follows Dana in considering fjords as drowned normal river valleys. Brigham[74] and Hull[75] seem to incline to the same view, the former speaking of "the common sense conclusion that they are river valleys made tidal by drowning"; but both recognize that fjords have been to some extent modified by glaciers. Hirt[76] in a review of "Das Fjord-Problem," Dinse[77] in a more elaborate study of "Die Fjordbildungen," and Grossman and Lomas[78] in a report on the Faroe Islands tend to assign to glaciers but a moderate rôle in modifying pre-existing valleys; while J. W. Tayler[79] and Fairchild[80] definitely reject the glacial theory of fjord formation, Fairchild specifically invoking coastal subsidence to account for the fjord embayments.

Among those students who admit that ice êrosion played an essential part in fashioning fjord valleys, there are a number who either expressly require coastal subsidence, or else tacitly assume that subsidence is necessary for the drowning of the glacial troughs. Robert Brown[81] writing on the "Formation of Fjords" in 1869 and 1871, required the combined action of glacial erosion and coastal subsidence. The same view is supported by Remmers[82] in his "Untersuchungen der Fjorde an der Küste von Maine," and by Güttner[83] in an essay on "Geographische

Homologien an den Küsten" published in 1895. Those writers assuming the necessity of subsidence without specifically discussing the point, include Penck[84] in his "Morphologie der Erdoberfläche," de Lapparent[85] in his "Traité de Géologie," Gallois[86] in his account of "Les Andes de Patagonie," Le Conte[87] in his "Elements of Geology," and Hobbs[88] in his "Earth Features and Their Meaning."

Formerly many observers were inclined to regard every fjord as either a rift valley formed by the dropping down of a narrow strip of the earth's crust between two parallel faults, or as a gaping chasm opened along a single fault. This tectonic theory of the origin of fjords, once much in vogue as an explanation for all valleys, is now generally regarded as obsolete. Statements of the tectonic theory in which ice is credited with a very minor rôle in clearing out crushed and broken rock left in the fault cleft, or in the moderate widening of an open chasm, will be found in a short paper by Gurlt[89], entitled "Über die Entstehungsweise der Fjorde," published in 1874, in Peschel's "Neue Probleme der vergleichenden Erdkunde als Versuch einer Morphologie der Erdoberfläche"[90], dated four years later; and in Kornerup's account of the fjords of southwest Greenland[91]. A more modern supporter of the tectonic origin of fjords is Steffen[92] in a short paper on "Der Baker-Fjord in West-Patagonien." But by far the most elaborate thesis in support of the tectonic theory is J. W. Gregory's recent book on "The Nature and Origin of Fjords"[93]. This serious attempt to rehabilitate a much-discredited theory of fjord origin contains extensive references to the literature of fjords, but frequently misinterprets the views held by the authors quoted. In a critical review of the book the present writer[94] has endeavored to show that Gregory's arguments are based upon a misconception of what the glacial theory of fjords implies, and upon an uncertain and variable interpretation of the tectonic theory.

Readers who wish to follow the discussion of the fjord problem further will be interested in an essay by Nordenskjöld[95] on "Topographisch-geologische Studien in Fjordgebieten" and in a shorter paper by Werth[96] entitled "Fjorde, Fjärde, und Föhrden." Both contain many references to the literature of the subject, and Werth's paper explains the differences between typical fjords, the allied forms in low rocky coasts like south-

PLATE XIX.

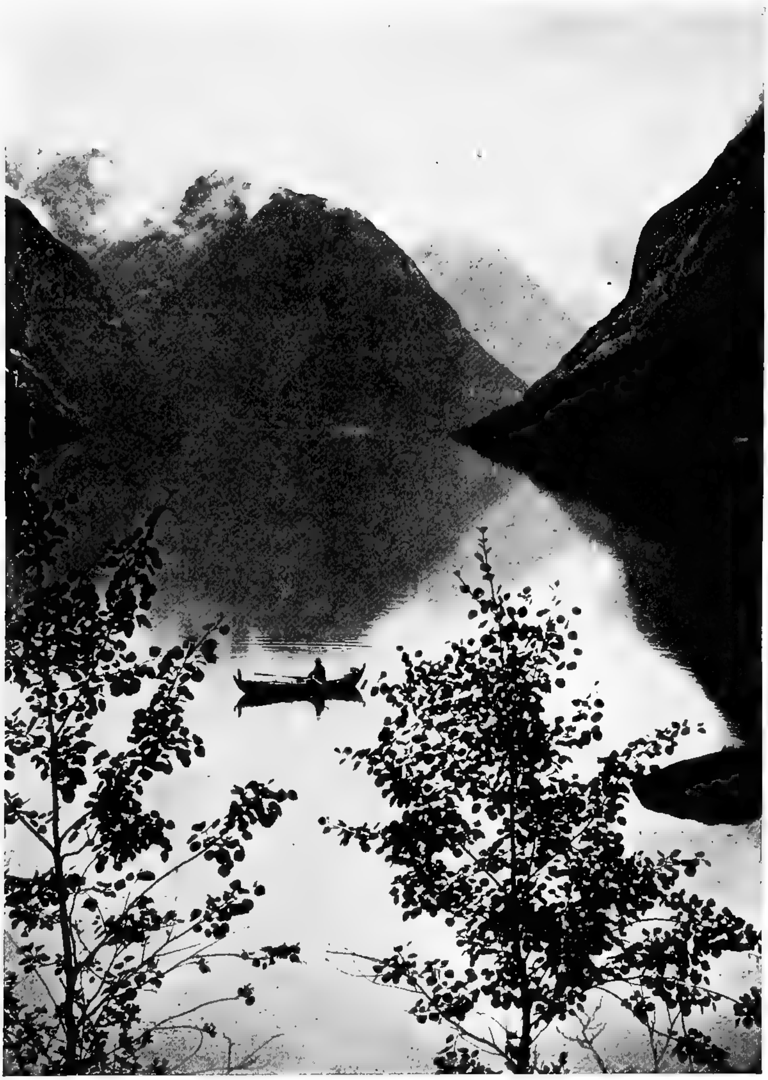


Photo by Underwood & Underwood.

Lake Loen, Norway, occupying a glacial trough and practically continuous with the upper part of Nord Fjord. Compare with similar topography shown on Plate XVIII.

western Sweden sometimes called *fiards* (Plate XX), and the *föhrden* of the Baltic shores of Denmark and Schleswig-Holstein, similar to fiards but lacking their rocky shores. The relations of these three sub-types of fjords are also considered by Penck[97], Dinse[98], and Hubbard[99]. An early paper by Ratzel[100] discusses at some length the essential characteristics of fjords.

Without, at this time, entering into any elaborate discussion of the several theories of fjord formation, it may be said that the interpretation which would regard fjords as partially submerged river valleys fails to account on any rational basis for the restriction of true fjords to glaciated high latitudes, for the identity in form between fjord-valleys and the glacial troughs of glaciated high altitudes, for the almost uniform violation of Playfair's law by tributary valleys which enter main fjords with discordant junctions, and for the occurrence of submerged fjord basins which, were the land to stand higher, would become lake basins not distinguishable from those of typical glacial troughs. Special pleading and strained reasoning have suggested a variety of possible explanations for each of these characteristic relationships, some of which might apply in one given instance, others in another. Glacial over-deepening of pre-existing river valleys alone offers a single explanation adequate to account at once for all of the specified relationships in all of the observed cases.

The tectonic theory of fjords is based on a misunderstanding of the significance of the known occurrence of fault-lines in certain fjords, and of the rectangular pattern of other fjords, which suggests an intersecting fault pattern. There can be little doubt but that crushed zones along faults, and infaulted strips of weak rock, have often determined the position and pattern of fjord-valleys. It is, however, an error of reasoning to jump to the conclusion that faults make fjords. As already noted, the glacial theory of fjord origin fully recognizes the fact that the pre-glacial valleys later transformed into fjords were often excavated along ancient fault-lines. Stream erosion naturally took advantage of the weak belts determined by faulting, forming fault-line valleys; but not until ice occupied these preglacial stream valleys and profoundly changed their shape and their depth, were the forms which we called fjords produced. To prove the presence of a fault-line through a fjord is, therefore, to prove nothing as to the tectonic origin of the fjord. The

PLATE XX.

Strömstad Harbor, Sweden, showing characteristics of a fiard coast.

tectonic theory, moreover, affords no rational explanation of the restriction of fjords to high latitudes, nor of the identity in form between fjord valleys and unsubmerged glacial troughs, between fjord-basins and trough lake-basins. In the glacial theory alone do all of the phenomena cited, including the relation of fjords to faults, find a logical interpretation.

It should be noted that the subsidence of the land, which was an essential element of the theory that fjords are drowned normal river valleys, has been specifically invoked or tacitly assumed by many of the supporters of the tectonic and glacial theories. Whatever may be said regarding the discarded tectonic theory, it is clearly an error of reasoning which would assume the necessity of land sinking in order to account for partial submergence of glacially over-deepened valleys. Glaciers in high latitudes reach the sea at the present time; and glaciers cannot cease to erode their channels until the ice is floated, which in turn cannot occur until the glacier has cut something like six-sevenths of its thickness below sealevel. Shaler was clearly right in stating that the over-deepened channel of such a glacier would be flooded by the sea as the ice melted, without any sinking of the land. It is important to remember, therefore, that there is no solid ground for the popular opinion that fjords are an indication of land subsidence.

II. Shorelines of Emergence. — The typical shoreline of emergence is the *coastal plain shoreline*, resulting from the emergence of a submarine or sublacustrine plain. Whatever may have been the initial inequalities of a given sea-bottom or lake-bottom, deposition of sediment will in time obliterate them. The resultant smooth bottom is protected from the action of those subaerial agents of erosion which normally give to the lands their remarkable variety of relief. Waves and currents tend to reduce irregularities of the subaqueous surface, not to produce them. We should, therefore, expect that most shorelines of emergence would be coastal plain shorelines; and this indeed seems to be the case.

It is conceivable that an irregular, dissected land mass might first be submerged, and then experience partial emergence before there was time for subaqueous processes to obliterate the irregularities. In such a case the shoreline would be classed as a shoreline of submergence, since all its chief characteristics are

determined by the earlier major movement of submergence and not by the later minor emergence. This is the case with the extremely irregular shoreline of Maine, which is frequently cited as a type example of the shoreline of submergence, notwithstanding a late uplift of the coast of moderate amount. Similarly the coastal plain of southern New Jersey and the coastal plain of Texas afford good examples of shorelines of emergence, although a later slight submergence has resulted in moderate embayment of the inner shorelines behind the offshore bars.

On theoretical grounds one might discuss other types of shorelines of emergence, as, for example, the shoreline which would be formed if an original submarine volcano were raised partially above sealevel by the upwarping of the ocean floor. Such discussion would not, however, materially add to our understanding of the principles of shoreline development, and may better be left to those who in the future encounter examples of such shorelines meriting special description.

FIG. 26. — Mississippi Delta. A typical lobate delta.

III. Neutral Shorelines. — While most of the world's shorelines have resulted from submergence of land areas or emergence of subaqueous surfaces, there remain important groups of shorelines whose essential characteristics depend on causes independent of either submergence or emergence. To this class of shorelines I propose to apply the term "neutral shorelines." Among others, the class will include the well-known (*a*) *delta shorelines* of variable form and extent. Where the current of a river's distributaries strongly predomi-

nate over shore currents and wave attack, the delta shoreline will be of the "lobate" type, as in the case of the Mississippi delta (Fig. 26). If shore currents or possibly wave erosion, or both, have a marked effect in shaping most of the accumulating delta deposit, but the river along one principal channel continues to advance its mouth into the lake or sea, a "cuspate" delta shoreline like that of the Tiber (Fig. 27) will result. In case either shore current or wave attack sets a limit to delta growth, even at the mouths of distributaries, what I have termed an "arcuate" delta shoreline may be formed, of which type the Niger delta (Fig. 28) seems to furnish a good example. Intermediate stages between these several types, or combinations of two or more types in a single delta, are frequently encountered.

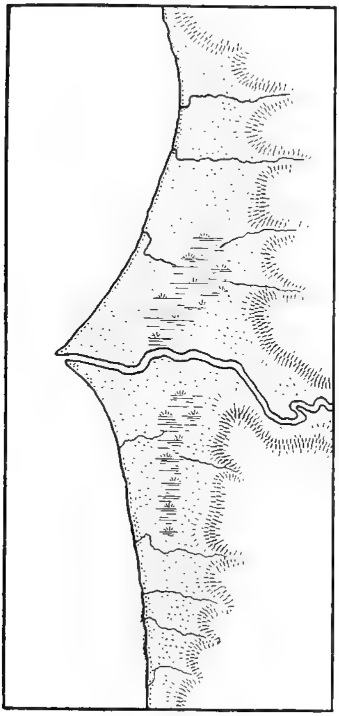


Fig. 27. — Delta of the Tiber. A cuspate delta.

Closely related to delta shorelines are (*b*) *alluvial plain shorelines*, and (*c*) *outwash plain shorelines*, formed where the broad alluvial slope at the base of a mountain range or the outwash plain in front of a glacier is built forward into a lake or the sea. On the landward side of such shorelines the topography is similar in many respects to that bordering the coastal plain shoreline, and the same may be true of the immediate offshore zone. Farther seaward the slope would normally become steeper, like the frontal slope of a delta. (*d*) *Volcano shorelines* of more or less circular pattern are formed where an active volcano, projecting above the water surface, builds its cone upward and outward by continued addition of ejected materials.

A very important group of neutral shorelines consists of (*e*) *coral reef shorelines*, formed when coral polyps build reefs upward

from a submarine floor or outward from the margins of any land area. Whatever the influence which past subsidence of the sea-bottom or elevation of the water surface may have exerted upon the particular forms assumed by coral reefs, does not affect the fact that the present shorelines of the reefs owe their existence to agencies which operate independently of such changes of level. The reef shorelines do not mark the contact of the water surface with pre-existing land areas or sea-bottoms, but with newly made land in process of formation at the present level. It would be out of place to enter here into a discussion of the much mooted coral reef problem. Those desiring to pursue this subject should consult the writings of Davis[101], and Daly[102], wherein discussions of the earlier work of Darwin, Dana, Murray and Agassiz will be found, together with copious references to the extensive literature published by other investigators of the problem. Vaughan[103] briefly discusses different theories of coral reef origin in a short paper published in 1916.

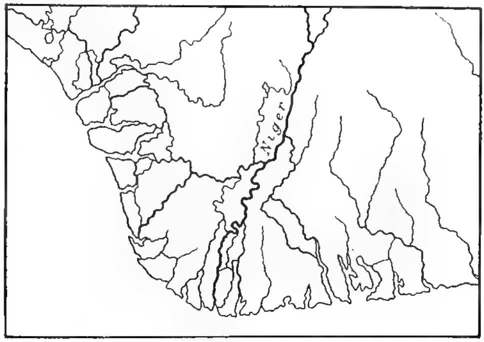


Fig. 28. — Niger Delta. The type of arcuate delta.

Another important group of shorelines belonging to the neutral class consists of (*f*) *fault shorelines.* The best discussion of this type of shorelines is contained in an excellent essay by Cotton,[104] to which we will recur on a later page. When the block on the downthrow side of a fault is so far depressed as to permit the waters of sea or lake to rest against the fault scarp, we have the typical fault shoreline (Fig. 29). Cotton describes shorelines of this type from near Wellington, and from other parts of New Zealand.

Earlier geological literature is full of references to shorelines or coasts supposed to result from faulting. Practically every irregular rocky coast has been explained as the ragged, broken edge of a land mass bordering a down-dropped segment of the

earth's crust. Descriptions of these coasts abound in such expressions as "fractured table-land bordering a foundered area," "the foundering of the adjacent ocean bed," "fractured margins of horsts," "the collapse of the basin of the Adriatic," and "shattered margin of the continent." Supported by the authority of men like Suess, the interpretation of irregular coasts as the

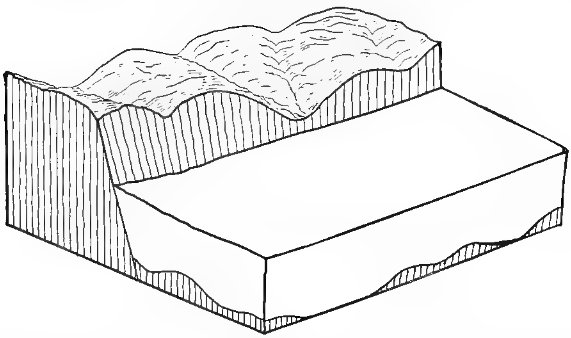


FIG. 29. — Initial stage of a fault shoreline. (Modified after Cotton.)

ragged edge of the land left standing when the adjacent area foundered beneath the sea, gained a currency, especially among German students[105], out of all proportion to its merits. It is now widely recognized that most, if not all, of these extremely irregular shores are better explained as shorelines of submergence, unrelated to faulting. Yet not a few writers of today, including occasionally a trained physiographer, show the influence of Suess' teaching by invoking the "shattering and foundering" theory for coasts like those of Greece, Dalmatia, and Norway. Fault shorelines exist; but so far as described by critical observers on the basis of competent evidence, they do not exhibit the irregular pattern of the coasts just mentioned. Shorelines of submergence, on the contrary, either of the ria or fjord type, show precisely those characteristics well displayed along the three coasts in question.

IV. Compound Shorelines are those which are prominently characterized by phenomena normally characteristic of at least two of the preceding classes. It frequently happens, for example, that oscillations in the level of land or sea leave a shoreline with a variety of features, some of which resulted from submergence, others from emergence. This is the case with the shoreline of

North Carolina (Fig. 30), which combines the drowned valleys of a shoreline of submergence with the offshore bar of a shoreline of emergence in such manner that it is difficult to decide which set of features is more prominent. We can therefore most properly speak of it as a compound shoreline.

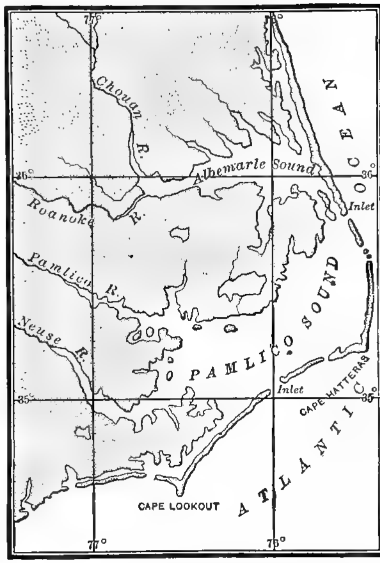


Fig. 30. — Coast of North Carolina, showing one type of compound shoreline.

In the formation of fault shorelines it may well happen that the block on the upthrow side of the fault is itself sufficiently depressed to permit drowning of the more deeply cut main valleys (Fig. 31). Such cases are reported from New Zealand by Cotton[106], and afford a very striking example of compound shorelines.

The term compound shoreline should be employed only when there is a very marked development of the features characteristic of two or more of the simpler classes of shorelines. Such

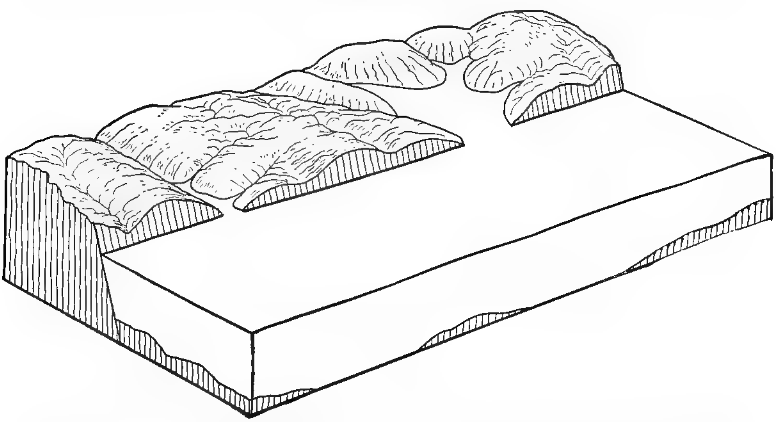


Fig. 31. — Compound shoreline due to faulting and partial submergence of upthrow block.

a shoreline as that of eastern Florida would be classed as a shoreline of emergence, notwithstanding the mild indications of submergence presented by the drowned valleys.

Stages of Shoreline Development. — The character of any shoreline depends, in the last instance, on the amount of work accomplished by marine agents upon the land against which the water surface comes to rest; or, in other words, upon the stage of shoreline development. Shorelines of submergence, shorelines of emergence, neutral shorelines, and compound shorelines of all varieties must therefore be further subdivided into groups according as they are in the initial, young, mature, or old stage of development, each group having its own peculiar characteristics. What these characteristics are will appear at some length in the following chapters.

RÉSUMÉ

We have outlined a terminology for the broader topographic features which characterize the margins of the land and sea. These features include four zones, known as the coast, shore, shore-face and offshore; three erosion forms, the cliff, bench, and abrasion platform; and three deposits called the beach, veneer, and continental terrace. All of these features are not invariably present, for we have already observed that the beach may be lacking, and it will appear in later chapters that one or more of the other features mentioned may fail to be developed in special cases. Shorelines have been classified into four main groups: shorelines of submergence, shorelines of emergence, neutral shorelines, and compound shorelines. The subdivisions of each class have been briefly considered, and some examples cited. We are now prepared to study the development of shores, by considering first the development of the shore profile, after which the shoreline itself will be treated.

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CHAPTER V

DEVELOPMENT OF THE SHORE PROFILE

SHORELINES OF SUBMERGENCE

Advance Summary. — In the development of any shoreline there are significant and systematic changes both in the profile and in the plan of the land margin. These changes take place in orderly sequence, and may best be described as the young, mature, and old stages of a cycle of shore development. We shall find, however, that the stages of shore profile development and the stages of shoreline development do not always keep pace with each other. A mature shoreline may have the shore profile of some of its parts still in the stage of youth; and it very commonly happens that a young shoreline has many points where the shore profile is mature. It will be desirable, therefore, to discuss the cycle of the shore profile first, and later to consider the cycle of the shoreline as a whole. In the present chapter the terms youth, maturity, and old age refer to stages of *profile* development only. The significant features of the profile involve all four of the zones adjacent to the shoreline, and when reference is made in the following paragraphs to the shore profile, it should be understood to include not only the shore proper, but the shoreface, offshore, and coast as well.

It has been deemed advisable to discuss first, and somewhat at length, profiles characteristic of the youth, maturity and old age of shorelines of submergence. Special attention is given to beach profiles and to their constantly shifting forms, matters of vital importance in many problems of marine engineering. A study of the ultimate stage of the shore profile leads logically to a consideration of the theory of marine planation. This theory is discussed fully and arguments presented to support the opinion that it merits a greater measure of confidence than most students of landforms are accustomed to accord it. The marine and fluvial cycles of erosion are correlated, their essential independence is emphasized, and their relative importance compared.

PLATE XXI.

Northwestern coast of France near Fécamp, showing youthful cliff profiles on a mature shoreline of submergence. High tide. Compare with Plate XXII.

In the final sections of the chapter the features peculiar to profiles across shorelines of emergence, neutral shorelines, and compound shorelines are given special treatment.

Initial Stage. — The initial profile at right angles to a shoreline of submergence normally indicates comparatively steep slopes descending rather abruptly into the water (a^1, Fig. 32). This is because submergence ordinarily permits the water level to come to rest against the hill-sides of the former land area. It is true that certain exceptions must be recognized. If the land area had been reduced to a peneplane surface before submergence, or if the form submerged were a young, undissected alluvial plain or other similar surface, then the initial profile will resemble that normally characteristic of a shoreline of emergence, and the history of development will be that appropriate for such a profile. Here we deal only with the more usual case, in which submergence permits the sea to come to rest against the irregular and comparatively pronounced hill slopes of a submature, mature, or late mature land mass.

Waves will at once attack the land at this new level, their vertical zone of activity extending from a short distance below sea-level to a short distance above; because, as we have already seen, the forward dashing crests of storm waves rise some feet above mean water level, while the vigor of wave activity dies out very rapidly below it. Although the waves do carry on a milder erosive activity at greater depths, the attack near the surface level is so much more vigorous that it is fair to liken the sea to a horizontal saw which cuts laterally into the land, the blade of the saw having a thick-

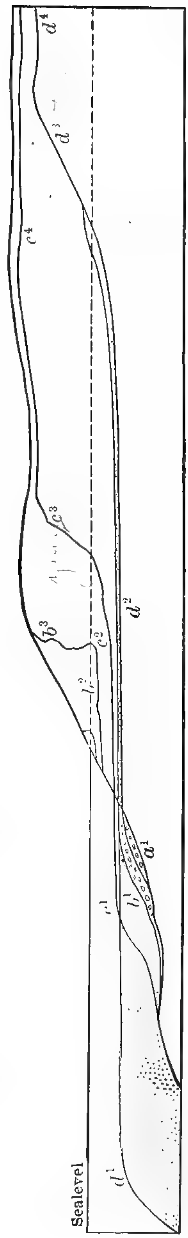
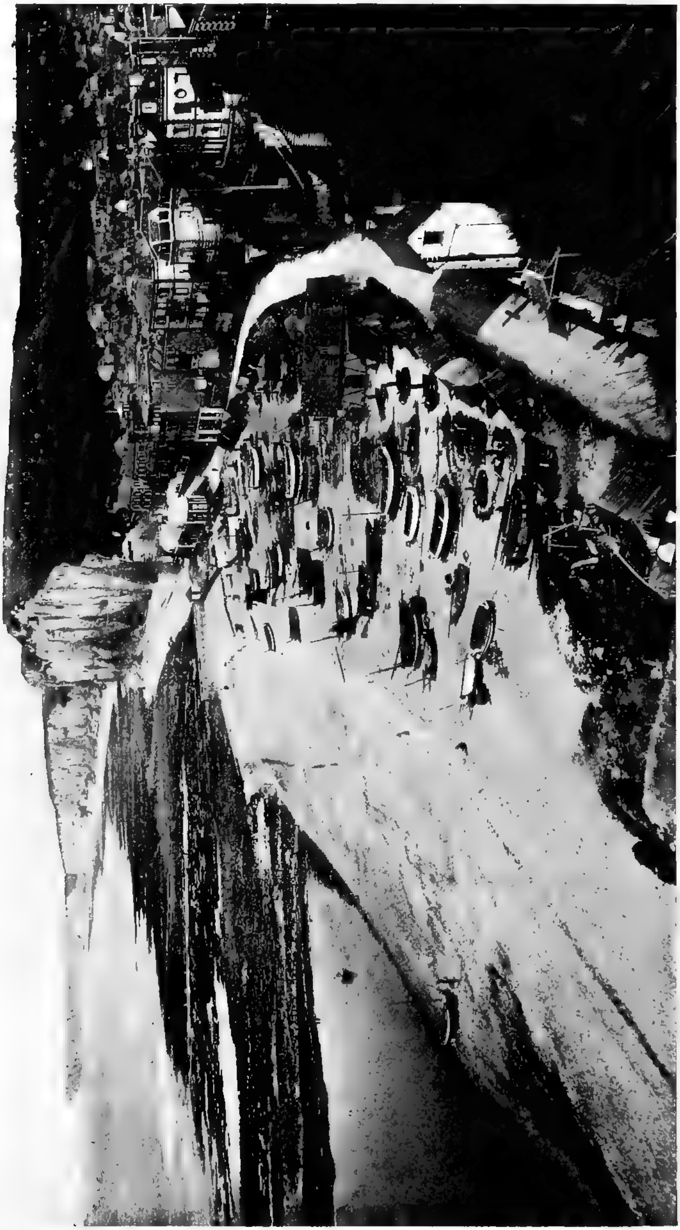


Fig. 32. — Stages in the development of the shore profile of a shoreline of submergence.

PLATE XXII.



Same view as Plate XXI, but at low tide, showing marine cliff, wave-cut bench (dark band), and narrow beach (light band) at base of cliff.

ness which extends from a few feet above to a few feet below sea level and being armed with breaking waves for teeth. Wave erosion soon cuts a notch in the edge of the sloping land, and thus destroys the initial profile. The coarse débris resulting from this erosion descends the underwater slope until it comes to rest as a submarine talus, where the water is deep enough to render wave agitation mild and the slope is gentle enough to require much agitation for the ready removal of coarse material.

Young Stage. — Continued wave erosion soon pushes the notch so far into the land that the unsupported overhanging rock falls down under the influence of the forces of weathering, including the action of gravity and rain wash on the face of the slope. This produces the wave-cut cliff (b^3), in front of which is the wave-cut rock terrace called the bench (b^2). The eroded débris will be added to the submarine talus (b^1) if the shoreface slope is steep enough and the water deep enough, except such part as is ground sufficiently fine to be widely distributed over the sea-bottom far offshore. If the water is shallow or the slope gentle, a shoreface terrace may be formed at this time.

In the cycle of stream development, the longitudinal profile of the young stream is characterized by irregularities which had their origin in the initial roughness of the land over which the stream flowed, or in the unequal erosion of alternate belts of resistant and non-resistant rock. Material eroded from parts of the profile exposed to vigorous cutting are deposited in the depressions of the profile where deep water is found. So also in the young shore profile we have irregularities due to the initial roughness of the submerged land mass, as well as irregularities in both cliff and bench (b^3, b^2) due to unequal resistance of the rock masses which are being eroded; and the material torn from the zones exposed to attack are deposited in the depressions of the profile where deeper and quieter waters occur.

As Davis[1] has shown, we may press this analogy even further with profit. In the typical young stream the water movement is vigorous because the initial slope of the land is comparatively steep, permitting a high velocity; transporting and eroding power are both great, but the transporting power is far more than sufficient to remove all the products of direct erosion. As the stream cuts downward the valley walls are undermined, and weathering causes the higher portions to descend into the stream

channel. But even the addition of these products of weathering does not over-tax the transporting power of the stream, and all the débris is swept down-valley to be deposited in quieter water below. The quantity of the products of weathering is not large, for the reason that since the stream has not yet cut deeply into the land, the valley walls are not high and therefore do not expose any considerable area to the forces of weathering. Because the rate at which the valley walls retreat, due to the forces of weathering, is not great as compared with the rate of valley deepening due to stream erosion, the slope of the valley walls is steep and may be vertical or even over-hanging in places.

Turning our attention to the shore, we observe precisely analogous conditions. Along the typical young shoreline of submergence the wave action is vigorous, because the initial slope of the coast is comparatively steep, permitting large waves to reach the land; transporting power and eroding power are both great, but the transporting power is far more than sufficient to remove from the base of the cliff all the products of direct erosion. As the waves cut inward the cliff is undermined, and weathering causes the higher portions to descend upon the marine bench. But even the addition of these products of weathering does not over-tax the transporting power of the wave currents, and all the débris is swept seaward to be deposited in the quieter deep water. The quantity of the products of weathering is not large, for the reason that since the waves have not yet cut far into the land, the marine cliff is not high and therefore does not expose any considerable area to the forces of weathering. Because the rate at which the face of the cliff retreats, due to the forces of weathering, is not great as compared with the rate of backward cutting due to wave erosion, the slope of the cliff face is steep, and may be vertical or even over-hanging in places.

A later stage of the youth of the shore profile shows some significant changes. As the waves cut farther into the land their power decreases because they must traverse greater and greater stretches of shallow water over the broadening marine bench; just as the stream which cuts deeper into a land mass suffers loss of erosive power because the water must flow more sluggishly on gentler and gentler gradients. But the loss of wave power comes at a time when the work to be done is increasing, for the increased height of the cliff enables the forces of weathering to

Marine cliff cut in glacial drift near Plymouth on the Massachusetts coast. The ordinary high-tide shoreline is marked by a line of débris.

cast a larger amount of débris upon the marine bench below; just as the higher valley walls of a deepening stream shed more waste into the channel at the very time the stream current is becoming more sluggish because of the decreased gradient. In both cases the work to be done increases as the power to do work decreases. A larger proportion of the weakening wave power must be consumed in transporting the increased amount of débris to deep water and in grinding the débris finer during the process of transportation, with the result that the base of the cliff is less and less vigorously pushed inland; just as a larger proportion of the weakening stream power must be used up in transporting the larger volumes of waste down-valley with the result that valley deepening is still further diminished. In the case of wave action, weathering now has the opportunity to wear back the marine cliff to a more gentle slope (c^3), which corresponds with the more gentle slopes of the valley walls in the similar stage of stream development.

Other important changes remain to be noted. During the appreciable length of time required for the pushing back of the cliff, the upland surface has been weathered and eroded to a lower level (c^4). Weathering of the cliff face goes on rapidly enough to keep pace with the enfeebled wave cutting at the base of the cliff, so that there is no longer a prominent notch at the level of wave erosion. The accumulation of the débris swept seaward from the marine bench by wave and possibly other currents has resulted in the formation of a shoreface terrace (c^1) whose top surface is delicately adjusted to continue the slight seaward inclination of the marine bench (c^2).

Still more important is the fact that the marine bench maintains its seaward inclination, and is therefore lower at its outer margin than it was at that same locality in an earlier stage of development. Thus the bench at c^2 has been lowered from the position b^2. There should be no difficulty in understanding this important change, and its causes and consequences. Waves continue to traverse the marine bench, and as the depth of the bench is not yet great enough to place it beyond the reach of wave action, it must suffer some erosion. The very fact that waves are weakened as they cross the bench toward the cliff proves that they have lost energy by expending it on the bottom. The débris weathered from the face of the cliff and eroded from

Marine cliff cut in sand and clays of the coastal plain near Beaufort, North Carolina. In the foreground is the wave-cut marine bench only partially covered by a thin beach deposit of shell fragments and other débris.

its base is dragged across the marine bench by wave currents, possibly aided by other currents, to be built into the shoreface terrace or moved into deeper water; and the long-continued action of this "marine sandpaper" must grind the surface of the bench ever lower and lower. As the outer part of the bench has been made longest and therefore exposed to continuous abrasion for the longest time, it is worn lower than the parts further landward. Thus the bench keeps its seaward inclination.

The effects of the seaward inclination of the marine bench are all-important. We have seen that waves tend to break when the depth of the water equals the height of the wave; hence the deeper the water the larger the waves which can traverse it. Progressive lowering of the marine bench therefore means the continuous admission of large waves farther and farther across its surface. Were it not for this lowering, a shallow, horizontal bench would greatly reduce the size and power of the waves which reached the cliff. While this would not completely stop cliff erosion, as has sometimes been assumed, it would enormously retard it. The seaward inclination of the bench greatly facilitates the removal of débris into deep water; for as we found from our study of wave action, if oscillatory waves produce equal impulses alternately landward and seaward, débris on an inclined sea-bottom must travel down the slope, whereas on a horizontal bottom it might remain in one place indefinitely. Effective removal of débris prevents it from protecting the cliff, and permits the waves to devote a greater proportion of their energy to cliff erosion. Thus in a second way the progressive lowering of the marine bench in such manner as to produce an inclined surface, greatly facilitates the recession of the shoreline under wave attack. A third important effect of the inclined bench is to raise the level of effective wave attack at the base of the cliff. We have observed in preceding chapters that winds blowing toward a steep coast with deep water offshore do not raise the water level appreciably, but that where the water is shallow its entire mass may receive a landward motion and thus pile up against the shore; that both oscillatory waves and waves of translation coming onshore raise the water level, waves of translation most effectively; that oscillatory waves change to waves of translation on a gradually shelving bottom; and finally, that tidal and other currents moving in upon such an inclined

PLATE XXV.

Marine cliff cut in sand dunes on the shore of Cape Canaveral, Florida.

slope raise the water level more effectively than when they impinge upon a steep slope which descends rapidly to deep water. All of these factors co-operate to raise the level of wave attack, especially during storms, to a slightly higher position as the shoreline is pushed inland. On the other hand, the development of strong waves of translation on the shallowing bottom during storms may move débris landward temporarily, thereby delaying its removal from the marine bench into deep water, and so retarding cliff recession for a time.

The notch at the base of the marine cliff is a measure of the ratio between wave erosion and weathering (including the action of gravity). When wave erosion is much the more vigorous, a pronounced notch occurs; when erosion exceeds weathering but slightly, the notch is only faintly developed, and if weathering is able to keep pace with erosion there will be no notch. Unconsolidated materials are quickly pulled down upon the marine bench by the action of gravity, which may be regarded as a very important element of weathering, since it is most efficient in promoting the disintegration of rock masses. As a result, erosion cannot gain on weathering sufficiently to produce a notch in sand cliffs and other unconsolidated material, even in the earliest stages of cliff erosion; whereas rocky coasts may possess good notches in early youth, faint notches in late youth, and none in maturity when weathering and erosion are delicately balanced.

On tidal shores, especially where the range of the tides is great, account must be taken of the varying water level. In the initial stage the vertical extent of the notch may be increased because of wave erosion throughout the whole extent of the tidal range. But early in youth it will be found that the notch is developed at the high tide level. Larger and more vigorous waves reach the coast in the deeper water of high tide, and the cliff is pushed in more vigorously at that higher level. The waves at low tide are left to expend their force on the shelving marine bench, and thus to assist in deepening its seaward portion. The general relations of the different topographic elements along the shore are not greatly different from those which would obtain if the high tide level were the mean water level of a tideless sea. Some minor differences will be noted as occasion demands.

Mature Stage. — The essential feature of maturity in the development of the shore profile is a condition of approximate

equilibrium between erosion, weathering, and transportation. In other words, the profile of maturity is, as in the case of the mature stream, a *profile of equilibrium*[2]. During youth the power of the waves to do work is far in excess of the work to be done. But as the development progresses the work to be done constantly increases, while the power to do work ever diminishes. There must come a time when the two are nicely balanced and equilibrium is established. This time ushers in the stage of maturity.

The essential nature of the shore profile of equilibrium may be better appreciated from an inspection of the accompanying diagram (Fig. 33). Where the cliff profile is steep (*c*) and much débris is shed into the water, the waves require a comparatively steep subaqueous slope in order that with the effective aid of gravity they may be able to remove the large amount of débris

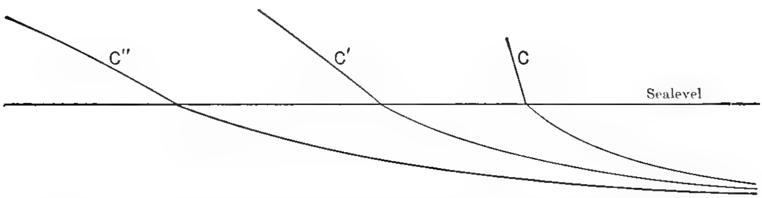


FIG. 33. — Successive profiles of equilibrium on a retrograding shore.

offered to them. With cliffs of progressively decreasing steepness (*c′* and *c″*), more gently inclined subaqueous slopes will permit that nice balance between the amount of work required to remove the diminished quantity of débris and the ability of the waves to do removal work, which we call "equilibrium." The subaqueous profile is steepest near the land where the débris is coarsest and most abundant; and progressively more gentle farther seaward where the débris has been ground finer and reduced in volume by the removal of part in suspension. At every point the slope is precisely of the steepness required to enable the amount of wave energy there developed to dispose of the volume and size of débris there in transit. Examples of actual profiles of equilibrium are shown in Figure 34.

Let us imagine that the profile d^1–d^3 (Fig. 35) is the shore profile of equilibrium, and verify the condition of equilibrium by

noting the consequences which must arise if we disturb any part of that profile. By assumption the erosion at the base of the cliff is just sufficient to supply the amount of débris which, added to the material contributed by the weathering of the cliff face, will provide the wave currents with the exact amount of material they can transport across the marine bench and shoreface

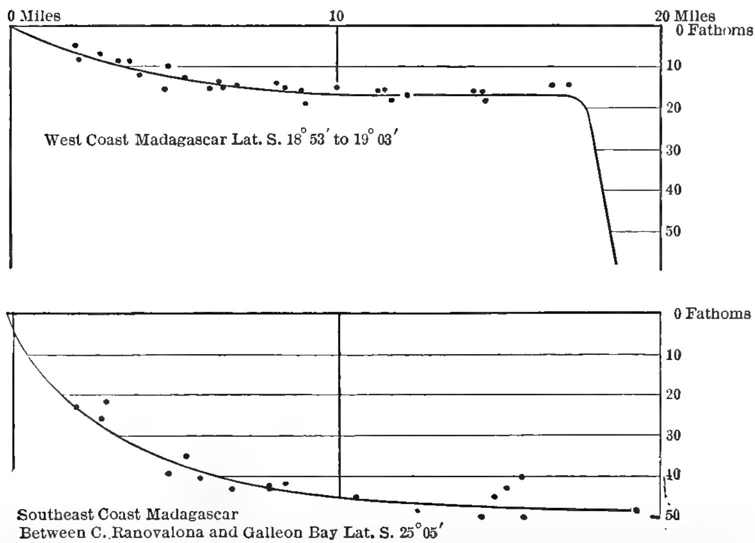


FIG. 34. — Profiles of equilibrium off the Madagascar coast as plotted from charts by Barrell. Note the striking difference between the profile of the protected west coast and that of the exposed southeast coast.

terrace to the front slope of the latter. Now, if we imagine the cliff (d^3) to weather more rapidly for any cause, this will mean an added accumulation of débris at the base of the cliff. The waves will have more material to transport and therefore less energy left to expend in erosion. Hence the base of the cliff is pushed back less rapidly than normally. But since the top of the cliff has weathered back more rapidly than usual, the ultimate result is a gentler slope for the cliff face. On the gentler slope weathering proceeds less rapidly than formerly, until the waves get rid of their excess burden and renew their erosion at the base of the cliff, thereby steepening it until weathering is once more normally adjusted to the other forces and equilibrium

is re-established. In a similar manner, if we disturb the equilibrium by increasing the wave erosion, this will mean more eroded material and products of weathering to be transported, wave currents will be overburdened, débris will accumulate unduly on the marine bench, thereby shallowing the water and decreasing the size of the waves which can reach the cliff base, thus reducing erosion until equilibrium is again restored. Increase of transporting power would sweep the marine bench clean and allow waves to deepen it more effectively, thereby admitting larger waves to the cliff base to produce greater erosion, and so increasing the material to be moved until the transporting power of the waves was again balanced by the amount of material requiring removal.

As in a mature river the equilibrium is never absolutely perfect, but rather an ideal condition which the stream ever strives to attain and does succeed in approximating very closely; so at the shore, where the variation in wave attack is far more irregular than stream volume and velocity, the equilibrium of maturity is only approximate. Each set of waves endeavors to establish a profile of equilibrium suited to its own needs, but seldom succeeds before another set of waves begins working toward a somewhat different profile. Fortunately, the small waves work so slowly as to effect no profound changes between times of vigorous wave action, while the attack of storm waves at a given point is sufficiently similar at different times to produce similar effects. There is, therefore, a certain characteristic profile of equilibrium for a given locality, notwithstanding the fact that

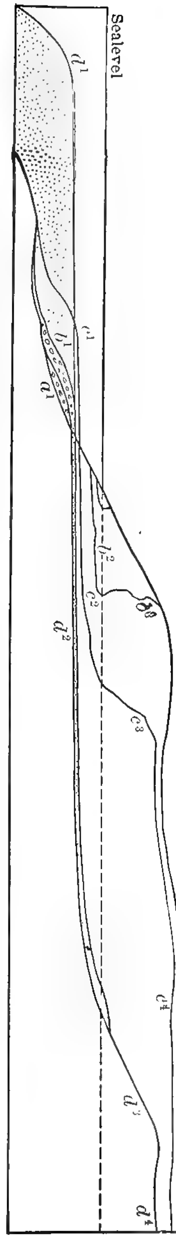


FIG. 35. — Stages in the development of the shore profile of a shoreline of submergence.

minor variations in the forces there at work will produce local changes which tend to confuse the student of shore forms. Let us first note the broader features of the mature profile, and then consider some of the more variable minor features.

In Figure 35 the profile d^1–d^4 is that of maturity. Comparing it with c^1–c^4, the profile of late youth, we note certain significant differences. The cliff d^3 has weathered back to a more gentle slope than in c^3, because wave attack is more feeble when the waves must traverse a broader marine bench which is encumbered, as we shall see, by more or less débris. As erosion carries the cliff farther and farther inland it will from time to time occupy positions on the landward-sloping sides of hills, in which positions the cliff decreases in height as it advances into the land (Plate VII). The presence of such cliffs of decreasing altitude along a coast implies considerable wave erosion in the past. The upland (d^4) has worn down to a lower level during the time required for the cliff to retreat from c^3 to d^3. As should be expected, the marine bench has likewise been worn lower at the same time that it has been extended inland; but it should be observed that although the cliff retreated twice as far from c^3 to d^3 as from b^3 to c^3, the outer part of the marine bench has not been lowered in proportion. This is because the waves act more feebly with increasing depth, and because the bench is more protected by débris than formerly. In consequence of the rapid decrease in wave power with increase in depth, the shoreward portion of the bench has a steeper slope than the portion in deeper water; or, in other words, the profile of the bench is faintly concave upward. A notable extension of the shoreface terrace (d^1) is apparent, the front of the terrace being convex upward. The compound profile of the bench and shoreface terrace combined is therefore roughly sigmoid, faintly concave upward near the landward end and convex at the seaward end.

The most important feature of maturity is the accumulation of débris on the marine bench to form a beach. During youth the vigorous wave action sweeps the products of weathering and erosion into deep water so rapidly that there may be no conspicuous deposits of waste on the bench most of the time. In maturity, however, the journey from the base of the cliff to deep water is so long, and wave action over much of the distance is so moderate, that any given moment may witness a considerable

quantity of débris in transit across the bench. Under normal conditions this material is not of appreciable depth, and its intermittent seaward movement serves to reduce the size of its component parts and to lower the level of the bench, by friction amongst the particles themselves and upon the bench surface. It is this beach deposit which undergoes the most sudden and repeated changes which characterize the shore and shoreface zones, and we may now turn our attention for a few moments to these changes and their causes.

The Beach. — In the first place, it must be borne in mind that the beach is merely a temporary deposit, slowly making its way to deeper water. If the various shore processes were perfectly uniform in their actions and always nicely adjusted to each other, the thickness of the deposit and its surface profile would remain essentially the same, while the component particles in the deposit would constantly migrate seaward and be as constantly replaced by new material weathered and eroded from the marine cliff and bench. But the forces are variable, both in character and intensity. Oscillatory waves may be replaced by waves of translation at irregular intervals; the undertow varies in volume and velocity and is modified by other currents; waves vary in size from day to day, and the storm waves of one season are more powerful than those of another season. All of these changes, and others that might be enumerated, disturb the equilibrium which would otherwise exist, and the beach deposit responds quickly to these disturbances. At one time wave erosion at the base of the cliff supplies material faster than it can be transported, and the beach deposit accumulates to a greater depth than usual. At another time waves fail to reach the cliff base for a long period, and the beach wastes away because the loss it suffers from continual attrition and removal under the influence of small waves is not made good by new supplies of débris. Again, waves of translation drive in much material from the shoreface terrace and even from the deeper water beyond, piling it upon the normal beach deposit and thereby greatly augmenting its thickness. Or storm waves accompanied by a vigorous undertow may sweep the entire beach from the marine bench, leaving the bare, solid rock exposed over extensive areas.

The factors involved in shore processes are so numerous, and their variations are so difficult to trace, that it is often im-

possible to ascertain just what disturbance of former conditions is responsible for a given change in the beach. As Hunt[3] has said: "A beach may resist the sea for years, yet in a few hours it may be stripped bare to the solid rock. Shells may be covering the bottom a mile offshore, undisturbed by onshore gales; a storm, with wind and waves apparently much the same as usual, may sweep them all onshore. One beach will be invariably kept clear of shells which will be found offshore, while another beach will have a constant supply, and for no obvious reason."

We may gain some appreciation of the extent of the above-mentioned changes in the beach deposit from the published reports of competent observers. Reference has already been made in an earlier chapter to the shingle and chalk ballast driven in upon the beaches between Tyne and Hartlepool, England, from points 7 to 10 miles offshore.[4] Along the coast of Algeria the waves cast large quantities of sand upon the beach, burying the roadway along the shore for considerable distances, a phenomenon well described by Fischer.[5] At one point on the Irish coast, according to Kinahan,[6] a beach 200 yards wide was built in front of a marine cliff during the spring of 1876, at a point where there was deep water the previous winter. The presence of a beach deposit along a shore for much of the time is apt to give one a false idea of its depth and stability. Thus many visitors to the beaches of the Atlantic coast find it difficult to realize that a single storm will often strip bare the underlying rock or expose buried peat deposits at places where they never see anything but an apparently inexhaustible store of beach sand. Hunt[7] refers to a case in which the eastern half of the shore at Blackpool, near Dartmouth, England, was stripped bare of its beach sands, for the only time in twenty-five years so far as was known. The bathing beach at Babbicombe was, according to this same author, so completely removed by a single storm that the place looked "as unlike a bathing-cove as any place can be." The southern coast of England is well known for its extensive beach deposits; yet Godwen Austen[8] writes: "I have seen, at one time or another, nearly every portion of our south coast in the condition of bare rock without sand or shingle. . . Bars, sand- and shingle-banks . . . are all subject to change of form and to removal, but they speedily collect again."

During the first few hours of a gale enough material may be removed from the shoreface to deepen the water there from 5 to 10 feet, especially where a sea wall helps to concentrate the wave energy along a narrow zone.[9] Along the Chesil Bank between Abbotsbury and Portland, Coode[10] estimated that a single storm removed 3,763,300 tons of shingle from the beach; and during another storm 4,500,000 tons of the shingle were scoured out, three-fourths of which was moved back after the gale ceased.[11] Pendleton[12] states that the shoreline of the beach along the southern coast of Long Island has varied temporarily back and forth 200 to 300 feet, due to storms.

Beach Profile of Equilibrium. — During all the temporary changes referred to above, the profile of equilibrium is maintained in as great perfection as the rapidly varying conditions will permit. Whether developed on the rock bench or on a thick overlying beach, the profile is concave upward. The concavity continues, with increasingly steep slope, above the normal water level, because the swash of the waves sweeps débris up the beach and deposits it in such manner as to maintain the necessary equilibrium between the onshore and offshore forces. Near the water line both the swash and the backwash of the waves have large volume and high velocity, and débris is swept back and forth on a fairly gentle slope. Farther up the beach the swash suffers loss of velocity because of increasing friction and the constant downward pull of gravity; and loss of volume because much water sinks out of sight into the interstices between the beach pebbles and sand. Consequently débris is deposited at the higher level of the beach and the backwash is too weak to return it to the sea. But the very act of deposition steepens the upper part of the slope, thereby increasing the effectiveness of the pull of gravity upon the débris, so that a small backwash can the more readily carry material back down the slope. Equilibrium is attained when the slope is so steep that the backwash aided by gravity can just return all the material which the larger swash can drive upward against the pull of gravity. In general, it may be said that in maturity the beach profile both in the shore and shoreface zones, is either nicely adjusted to the conditions imposed by a set of waves which have been operating for some time, or is rapidly undergoing adjustment to a new set of waves which differ from those previously operating. Let us note some of the changes

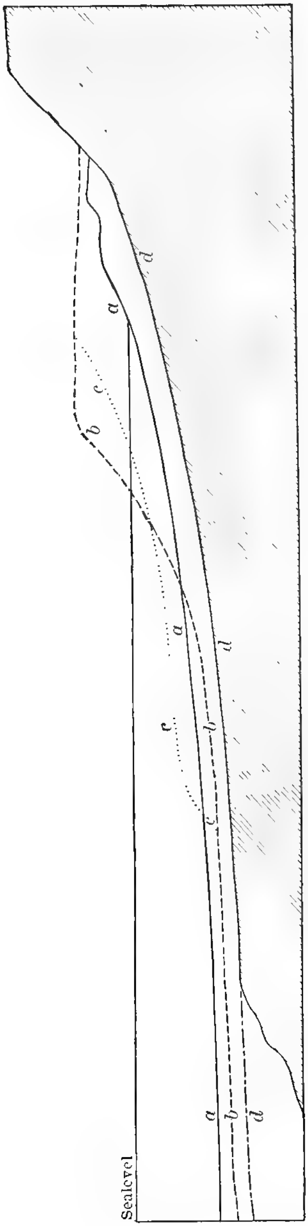


FIG. 36. — Variations in the beach profile of equilibrium due to variations in the different shore forces.

in the beach profile which result from these adjustments to varying conditions.

Imagine a mature shore profile (*aaa,* Fig. 36) in which a thin beach deposit covers the marine bench and is continued seaward by the shoreface terrace. First let us suppose that a series of oscillatory waves encountering the seaward edge of the terrace are partially transformed into waves of translation. The waves of translation will then drive the bottom débris landward and bank it up against the base of the cliff, building the beach deposit forward and making its front of such steepness that gravity plus undertow will just balance the tendency of the shoreward component to carry material up the slope. But the taking of material from the bottom deepens the water, and deepening water is more and more unfavorable to the development of waves of translation. The waves retain more of their oscillatory character than formerly; and with the more sudden descent into deep water in front of the new deposit the undertow becomes more effective, finally overcoming further efforts toward landward transportation. We will then have the profile *bbb,* which is the profile of equilibrium under the new conditions.

Now let us imagine that this new profile is subjected to the

action of smaller oscillatory waves, in which the offshore components (backward oscillation + gravity on the steep slope + undertow) are in excess of the shoreward components. Material will be eroded from the upper part of the deposit and carried seaward. But the undertow associated with these smaller waves does not possess a great transporting power. Consequently much of the seaward moving débris will be quickly dropped, thus building up the bottom to a higher level and decreasing the water depth. The effect of this is to restrict the undertow within smaller limits and so to increase its velocity until it is able to transport all the material eroded by the waves. Thus equilibrium between the various factors is once more perfected. Since the load of moving débris is in equilibrium with the transporting currents at a higher level than before, a new shoreface terrace *ccc* builds forward over the former bench, and possibly over the older terrace.

Finally, let us imagine that a series of great storm waves, accompanied by a vigorous undertow, attacks the shore under consideration. Erosive power is great enough to cut into the beach deposit and remove it, and possibly to attack the cliff itself; and the seaward currents along the bottom are more than strong enough, at the higher level of the profile *cc*, to transport all débris. They therefore erode the bottom, deepening the water, and thus decreasing their velocity until they are just able to transport the material delivered to them. It may well be that this new equilibrium is not reached until the marine bench is swept clean and the profile of bench and terrace reduced to the line *ddd*.

Other changes in the profile of the beach must result from other variations in the on- and offshore forces. An offshore wind may cause a landward bottom current, as we have already seen, and this, aided by wave agitation, builds the beach forward until the front slope is so steep and the water so deep that equilibrium is restored. An onshore wind may develop such a vigorous undertow that the bench will be stripped of much of its deposits before equilibrium is again reached. This would explain Coode's[1] observation that after offshore winds the slope of a shingle beach is 1 in $3\frac{1}{2}$ or 4; whereas after heavy onshore winds the slope is only 1 in 9 or $9\frac{1}{2}$. If the supply of waste from the cliff is stopped for any reason, the beach will be removed and the bench lowered

to a new profile. On the other hand, if a change in the character of cliff material should result in a more rapid supply of débris, the seaward currents will be too weak to transport all the débris until deposition has shallowed the water and thereby increased the current velocity; or, as Fenneman[14] has expressed it: "If the supply of material be suddenly increased, a smaller shelf will grow from shore on the surface of the older, for the reason that the new load, being greater, is in equilibrium with the currents at a higher level than before." In all of these and other similar changes, the profile of equilibrium is either maintained or quickly restored.

From what has been said in the preceding paragraphs, it is evident that man has the power to retard cliff erosion if he can deposit a sufficient amount of débris upon the shore to overload the waves and cause them to establish a profile of equilibrium which does not touch the bare rock of bench or cliff. On the other hand, man may accelerate cliff erosion by removing sand or shingle from the beach, thereby causing the waves to expend their excess energy in retrograding the shoreline until a new profile of equilibrium is established. Legal authorities have taken cognizance of this latter possibility in a number of cases. Thus the British Board of Trade has repeatedly prohibited the removal of beach material from shores where it was clear that such removal would be injurious to the coast. In an action brought by the Attorney General against a certain lord who asserted his right to remove shingle from his own shores, it was held that it was the duty of the Crown to protect the realms from inroads of the sea by maintaining the beaches in their natural condition; and an injunction was granted restraining any further removal.[15] When the removal of shingle from the beach at Spurn Point, England, for road mending and concrete, was stopped, the erosion of the cliffs diminished one-half.[16] On the Prussian shores the taking of stones from the beach is "polizeilich verboten."

I have dwelt at some length upon the local and temporary variations in the profile of equilibrium, in order to make clear their essential unimportance so far as the whole cycle of shore profile development is concerned. This is the more necessary because the true significance of these changes has not been as widely understood as one could wish were the case. Long ar-

ticles have been written, extended discussions have been carried on, and numerous erroneous laws of shoreline activity have been laid down, all based on observations of minor fluctuations in the shore profile of equilibrium. This has been unfortunate for the development of that part of the science of physiography relating to shorelines, for two reasons: It has concentrated attention on the less important details of shore activities, and caused a neglect of the broader and more fundamental aspects of coast erosion; and it has led to endless controversy regarding the conditions of wave erosion and deposition, and the relative importance of waves, winds, and tides in controlling the direction of débris migration along the coast.

The problem of longshore débris migration will be taken up in a later paragraph. Emphasis may here be laid upon the fact that the shore profile of equilibrium represents a condition of balance, not between two forces but between many forces. Whether a beach will be eroded or will have material added to it does not depend upon the number of waves which arrive per minute; nor upon whether the waves are groundswells or local wind waves; nor upon whether the waves strike the beach obliquely or at right angles; nor upon whether the wind blows with the waves or against them; nor upon whether the waves run with the tide or against it; nor upon whether the waves are of the oscillatory or translatory variety. Absolute rules regarding the behavior of beaches under each of the above conditions have been published, some of which have been quoted on previous pages. Yet all these rules are necessarily fallacious because they take no account of the fundamental fact that beach erosion or deposition must ultimately depend upon whether or not the profile is in equilibrium with the resultant of all the forces operating upon it. In a complex of forces, it is not permissible to pick out some one force and attempt to build theories upon its sole activity; for it may well happen that its effect may be overcome by the superior power of other forces associated with it. We can thus readily understand the fact that every "rule of thumb," relating to wave action on beaches, yet proposed has been vigorously assailed by men whose observations directly contradicted it. I shall hope to show in the pages which follow that the matters thus elaborately debated are of relatively small consequence, in view of the fact that the ultimate

tendency of all wave action is to erode the lands. The temporary variations in beach and bench profiles are insignificant incidents in the relentless advance of the waves into the heart of the continents.

Effect of Longshore Currents.—Thus far attention has been directed to the very temporary changes in the shore profile resulting from variations in the activity of on- and offshore forces. Let us now consider the effect of longshore current action upon the shore profile of maturity. In the first place, if the longshore action be in the nature of beach drifting it is evident that anything which locally stops that movement must force a readjustment of profiles on both sides of the obstruction. For there will be an undue accumulation of material on the near side of the obstruction, causing a prograding of the shore until the profile is steep enough to allow the offshore forces to dispose of the excess material. On the far side of the obstruction the shoreline will be retrograded, because the failure of the longshore supply of débris will leave the shore forces with an excess of energy which will be expended in erosion. It is for this reason that the erection of a pier or groin, extending out from a gravelly or sandy beach, is usually followed by an advance of the beach on one side and a cutting away of the beach on the other side of the structure.

If the longshore movement be in the nature of a more extensive current located some distance offshore, the results may be far more impressive. Imagine a shore in which the profile of equilibrium is established, and is being gradually pushed landward under wave attack, accompanied, of course, by the minor fluctuations in beach profiles which have been discussed above. Now let us suppose that a broad current of any type flows parallel with the shore, bringing with it much débris, a part of which is deposited in the offshore zone. Continued deposition shallows the water, thus favoring the development of waves of translation. As we have already seen, waves of this type tend to remove the deposited material from the bottom and drive it landward, adding it to the front of the beach. Normally, the effect of this action is to leave deeper water offshore, which is in turn unfavorable for the development of waves of translation. But in the case before us the longshore current continually shallows the bottom by deposition; hence waves of translation

may continually form, and constantly add material to the front of the beach. Just so long as the current aggrades (builds up) the seabottom offshore, the waves will prograde (build forward) the shore. Following Davis we may call any shore which is experiencing such a long-continued advance into the sea, a *prograding shore,* and distinguish it from the more usual retreating or *retrograding shore.*

The prograding of a shoreline may take place rapidly or slowly, and may continue for a few years, a few centuries, or many thousands of years. According to Marindin[17] the beach at Siasconsett, Nantucket Island, has advanced 255 meters between the years 1846 and 1890. A beach in front of a marine cliff at Nantasket, Massachusetts, has grown seaward 400 meters or more during a period estimated at one to three thousand years.[18] The shore of the Darss, in northern Germany, has been prograded 7000 or 8000 meters since about the year 2000 B. C.[00] and a somewhat greater length of time was probably required for the advance of Cape Canaveral a similar distance into the sea. It should be borne in mind, however, that these long-continued additions to the land, while far more important and significant than the minor fluctuations previously discussed, are themselves only temporary effects of longer duration, and that in a comparatively short fraction of the whole shoreline cycle they must be cut away. This point will be further considered on later pages.

Longshore currents which have a fairly high velocity but which bring little or no sediment to deposit, may help to keep the marine bench swept clean of débris, thus materially aiding the retrograding of the shoreline. It is probable that some of the localities where a broad marine bench is usually well exposed, as for example off some parts of the coast of Brittany, owe the exposure of the bench to the effective assistance which wind, tidal, or other currents lend to the normal on- and offshore processes.

There remain for consideration one or two minor features of the shore profile of maturity. The landward portion of the profile is apt to be complicated by a series of "storm beaches" or "storm terraces," representing the effects of waves of varying dimensions at different heights of the tide. Such backshore terraces often have a faint landward slope on their upper surfaces. This is due to the fact that overwash from the highest

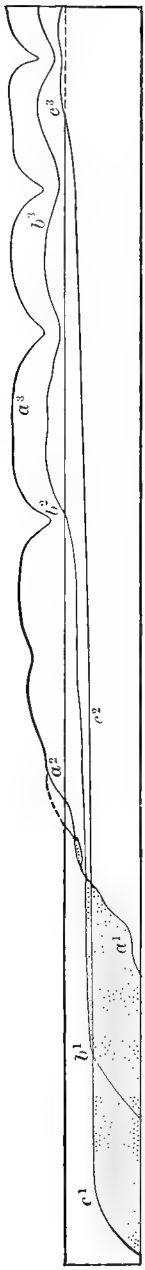


FIG. 37. — Mature and old shore profiles of a shoreline of submergence. On a much smaller scale than Fig. 35.

waves flows across the terrace, depositing a larger proportion of its load near the seaward edge, where its velocity is first checked and its volume is rapidly decreasing from loss of water sinking into the porous beach deposit. The front of each terrace may represent the upper part of a former beach profile of equilibrium; or it may be an erosion scarp if waves at lower levels have cut into the former profile, instead of merely depositing débris in front of and upon it. Beach cusps[20] may give the front of any backshore terrace a serrate plan.

The shoreface terrace has an upwardly convex profile at its seaward margin, as we have already noted; but the foot of the terrace may have a concave profile due to deposition from suspension.[21]

Old Stage. — In Figure 37 the profile $a^1a^2a^3$ represents a partially submerged land mass with a mature shore profile at a^2 where we see the cliff, bench, and terrace which are shown on a larger scale in Figure 35, $d^1d^2d^3$. On the scale of Figure 37 it is not practicable to represent the beach deposit on the bench, but its presence may be inferred. The profile $b^1b^2b^3$ is the profile of early old age, and $c^1c^2c^3$ the profile of advanced old age of this same shore. It will be noticed that in early old age the cliff (b^2) has a very faint slope, scarcely meriting the name "cliff," except in a technical sense; for wave erosion must proceed slowly when the waves have to traverse the vast expanse of shallow water over the wide abrasion platform which they themselves have cut, and when all débris must slowly be moved from the cliff to the edge of the terrace at b^1 before it reaches a final resting place. The marine bench is still to be found in front of the cliff; but it merges imperceptibly into the similar but much larger and more faintly inclined erosion surface which we have just referred

to as the abrasion platform. The continental terrace (b^1) is developed on a large scale; and the abrasion platform, normally covered with a thin marine veneer in excessively slow transit, is now so broad that, in combination with the continental terrace, it gives a very extensive continental shelf. One would scarcely expect all eroded débris to be transported to such a distance as to leave a continental shelf consisting wholly of an abrasion platform, although this appears to be Vogt's idea of the "kontinentale plattform" off the northern coast of Norway.[22] Erosion and weathering have reduced the former upland (a^3) to a series of broad valleys separated by subdued divides of moderate elevation (b^3).

In advanced old age the cliff (c^3) has been pushed much farther inland, and is so low and flat that it is almost imperceptible. The remaining land area has been worn down to a peneplane of faint relief. Abrasion platform (c^2) and continental terrace (c^1) are much broader than before. Waste is supplied so slowly from the land that the abrasion platform is gradually denuded of its veneer, and the vast extent of continental shelf may consist largely of bare rock on the landward side and sedimentary deposits on the deep-water side.

Wave Base. — The final stage of marine erosion will have been reached when the entire land mass is reduced to an ultimate abrasion platform surrounded on all sides by a continental terrace, the level of the platform being as far below the surface of the sea as wave erosion is effective. In other words, the cycle of marine denudation is completed when all the land is reduced to the baselevel of wave-erosion, just as the cycle of fluvial denudation is completed when all the land is reduced to the baselevel of stream erosion. The valuable term "wave base" was introduced by Gulliver[23] to denote the imaginary plane down to which wave action tends continually to reduce the lands; and since, as we have seen, the lower limit of effective wave work is probably reached at a depth of about 600 feet, we may tentatively consider wave base as an imaginary plane about 600 feet below the surface of the sea. A cycle of wave erosion ends, therefore, when all the land is reduced to a plane surface about 600 feet below sealevel.

A common error is to confuse *wave base* with *profile of equilibrium.* The gently sloping subaqueous terrace bordering lake

shores or the shores of a sea, and the submarine platforms of islands truncated by ocean waves, are frequently explained as the products of wave erosion down to wave base, when in fact they merely represent surfaces of equilibrium which are very slowly being reduced toward a wave base far below. The Platte River has established its profile of equilibrium at a level which is in places some thousands of feet above the sealevel, and to ordinary observation does not now appear to be cutting its valley any deeper. Yet no one would make the mistake of saying that the valley floor of this stream had been reduced to baselevel. It is no less erroneous to say that a subaqueous terrace on which the marine forces are now in equilibrium and which shows no evident indications of being cut deeper, has been reduced to wave base. The error is compounded when the false assumption that the terrace represents wave base is made the ground for the conclusion that wave action is not effective below a comparatively shallow depth. Equilibrium may be established at a shallow depth; from that level downward wave erosion proceeds more and more slowly, but none the less surely.

It might seem on first thought that no limit could be set to the depth of wave action, because theoretically waves of translation affect the water on the bottom as much as they do the surface layers, no matter what the water depth may be. We have already seen, however, that waves of translation are more apt to be formed in the shallow waters surrounding the lands, since conditions favorable to their development seldom exist in the deep sea. Another point of much importance in this connection is that waves of translation, when propagated into deep water, tend to change into oscillatory waves, as has been shown by Rankine.[24] It would seem to follow from this that the ordinary waves of translation found near the shores cannot be efficient agents in lowering the level of wave base, because they cease to exist as such when the water attains any considerable depth.

Gulliver[25] states that the abrasion platform "will not lie as far below the surface of the sea [in late old age of shore development] as it did in its maturity." Such a statement suggests that Gulliver confused the plane of denudation or abrasion platform with the submarine plain of deposition formed by the marine veneer laid down upon the platform. Even so, it is difficult to see how the marine veneer could become thicker in the late old

PLATE XXVI.

Photo by R. Lunn.

View near Rhuvaal, Islay, off the west coast of Scotland, showing former marine cliff and wave-cut bench, recently elevated above sealevel. At the water's edge the waves are cutting a new cliff and bench at the present level.

age of shore development, thereby shallowing the sea. There is a slight tendency in this direction at an earlier stage; but in late old age the supply of débris from the land is decreased and the abrasion platform must be denuded of its veneer, as already shown.

Validity of the Theory of a Marine Cycle. — Thus far I have for the most part assumed the theoretical possibility of extensive marine erosion to wave base, and have only incidentally referred to contrary opinions. It is only proper, however, that we fairly consider any objections to the theory of marine planation and determine whether they invalidate any of the conclusions reached above.

Wave-cut Benches. — One finds no reason to doubt that wave erosion has produced more or less plane surfaces of moderate breadth around the margins of certain lands. The very striking pre-glacial shore terrace (Plate XXVI) bordering the v [illegible] isles of Scotland is described by Wright[26] as an uplifted platform of marine erosion having a breadth of about half a mile in places, part of the breadth having been lost through later wave erosion at present sealevel. Lawson[27] has described uplifted wave-cut rock platforms on the coast of California having a maximum width of more than a mile. Comparatively rapid emergence of the land prevented long-continued wave attack at one horizon, with the result that the platforms constitute an extensive series of terraces (Fig. 38), the highest of which is over 1500 feet above sea level. There can be no doubt that had all this erosive work been performed at one horizon the resulting platform would have been much broader than any one of the existing wave-cut surfaces. Comparatively weak waves on Lake Michigan attacking shores of glacial drift have formed a terrace whose outer margin is approximately 60 feet below the lake surface, and which varies in breadth from 2 to 6 miles, with a maximum at one locality of 12 miles. Andrews[28] assumed that the entire breadth of the terrace was due to wave erosion; and proceeding on the further assumption that the rate of wave erosion is the same during all stages of terrace cutting, he used the breadth of the terrace and the present known rate of cliff retreat to establish a measurement of post-glacial time. Both his assumptions must be considered erroneous; but it seems probable that from one to several miles of the terrace breadth is wave-cut, even though a

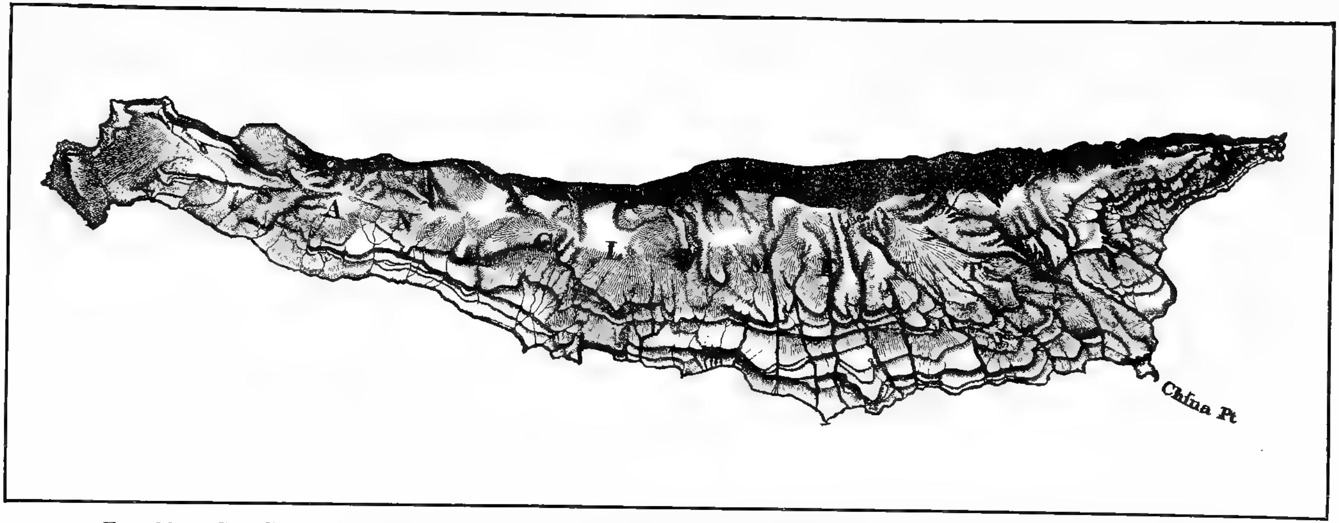


FIG. 38. — San Clemente Island off the coast of southern California, showing series of uplifted wave-cut terraces. From U. S. Coast Survey Chart No. 5100.

large part of it represents the effect of wave deposition. In a paper entitled "Fenomeni di abrasione sulle coste dei paesi dell' Atlante"[29] Fischer describes a submarine terrace bordering parts of the north coast of Africa having an outer margin approximately 100 to 200 meters below the surface of the Mediterranean, and a maximum breadth of at least 12 miles. The entire breadth of this terrace is regarded by Fischer as a marine abrasion platform; but it seems probable that the outer part of it is of constructional origin. Good photographic illustrations of its exposed landward margin, where it is an undoubted platform of marine abrasion, accompany the same author's report on "Küstenstudien und Reiseeindrücke aus Algerien,"[30] while a short description of the terrace occurs in an earlier paper on "Küstenstudien aus Nordafrika."[31] It is well known that certain volcanoes formed in the ocean have been reduced by wave erosion to submarine platforms within the space of a few y
and there are excellent reasons for believing that many of the more or less circular submarine platforms in the Pacific Ocean described by Wharton[33] and other writers, and more recently discussed by Daly[34] in connection with the glacial-control theory of coral reefs, represent volcanoes whose summits have been truncated by marine abrasion. Not a few of these platforms measure from 20 to 30 miles or more in diameter, but what portion of the whole represents marginal deposits of débris eroded from the center is unknown.

The great wave-cut platform ("strandfladen" of the Norwegians) fringing the west coast of Norway, best known through the studies of Reusch,[35] Richter,[36] Vogt[37] and Nansen,[38] has an average breadth of nearly 30 miles, and a maximum breadth of nearly 40 miles according to Vogt and Nansen, if we include the portion still submerged. Notwithstanding the doubt implied by Reusch, and clearly expressed by Hansen[39] and Nussbaum[40] regarding the essential marine origin of this topographic feature it is generally considered, and probably correctly so, one of the best examples of marine abrasion on a large scale yet discovered along our present coasts. Nansen[41] describes similar platforms of marine abrasion fringing the coasts of Siberia, Greenland and other land areas, none of which are so broad as the Norwegian case, although a breadth of nearly 20 miles is not unknown. The east coast of India, as described by Cushing,[42] consists

in part of a remarkably smooth, uplifted plane of marine denudation, above which rise numerous unconsumed remnants of quartzite, the bases of these former islands or stacks not infrequently being marked by sea-caves (Plates XXIX–XXXI). In places this wave-cut plane attains a breadth of about 50 miles.[43]

It seems highly probable that considerable portions of the continental shelves bordering certain shores represent platforms of marine abrasion. Nansen[44] is of the opinion that a great part of the continental shelf west of Norway is of this origin, and believes that between latitude 65° 10′ N. and 66° N. solid rock is present clear to the edge of the shelf. Figure 39 represents two of Nansen's sections for this region, in which the results of soundings are indicated. Notwithstanding the difficulty of determining the presence of solid rock by the sounding method, Nansen believes that the rocky ridge shown near the outer margin of the shelf is correctly indicated. If he is right we have here a plane of marine abrasion, including the rocky "coast platform" described above, exceeding 170 miles in maximum breadth.

It is true that Nansen doubts the power of waves to carve a broad and gently sloping platform on a simple coast; and he therefore assumes that even in the case of the narrower "coast platform" or "strandfladen" the coast was first deeply indented by fjords, and the platform later cut during glacial and interglacial periods by wave attack from both the ocean and the fjord waters, aided by subaerial denudation.[45] We must doubt the validity of the theoretical grounds on which he thus limits the power of waves in the open ocean; and must likewise doubt whether small waves in sheltered fjords, formed as they are on the surface of deep water and therefore unarmed with débris, and subject to reflection from nearly vertical rock walls without opportunity for erosion, could materially aid in the process of reducing a land mass to a submarine platform. Neither do his arguments in favor of the glacial age of the coast platform appear convincing[46]. But the facts presented by this author leave no room to doubt the existence along the west coast of Norway of a wave-carved platform which is certainly 50 to 75 miles broad, and possibly as much as 170 miles broad in some of its parts.

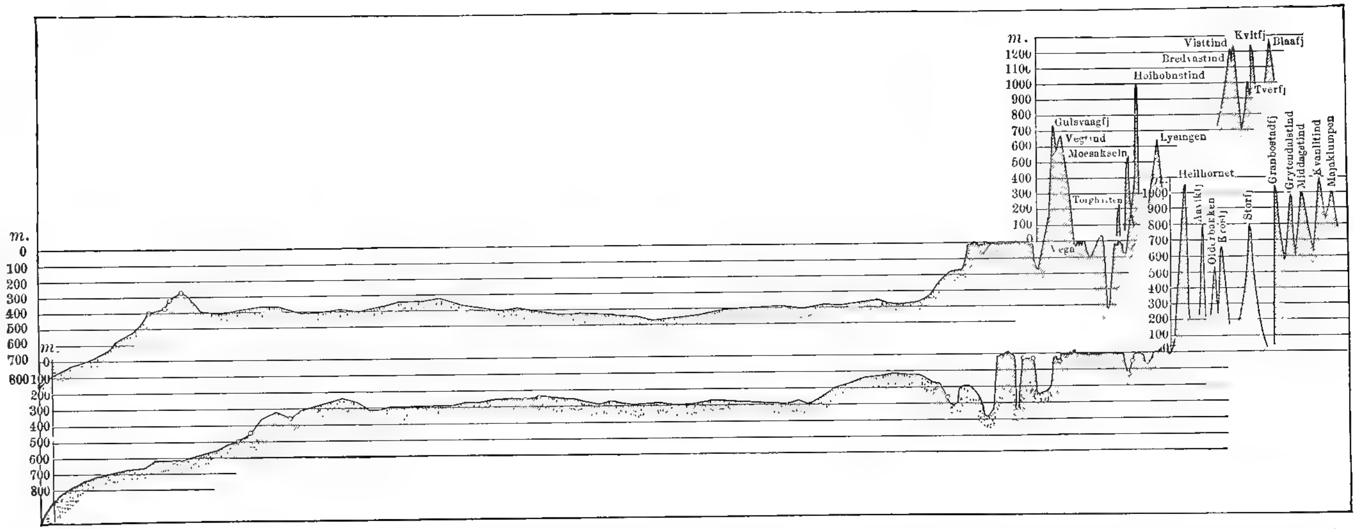


FIG. 39. — Cross-sections of the western coast and continental shelf of Norway. The wave-cut "coast platform" near sealevel (m.0) separates the mountainous highlands from the broad continental shelf. Cross-hatched zones indicate solid rock, dotted zones indicate sands, muds, etc. (Modified after Nansen.)

Elevated marine cliff near Oban, Scotland. The haystacks stand upon a wave-cut rock bench thinly covered with sand and gravel, now raised 25 feet above sealevel.

Theory of Marine Abrasion.—It is evident from the brief survey given above that planes of marine abrasion a great many miles in breadth are well-attested features of the earth's surface. But it is difficult or impossible to tell whether or not these planes were formed during a still-stand of the land or during a progressive submergence. There is a widespread idea that waves can cut into a still-standing land mass only to a very moderate extent before they will exhaust themselves on the shallow bench which they have carved. According to this interpretation, a marine cliff will be pushed inland by the waves for a short distance, and will then remain unchanged in position unless subsidence of the land mass deepens the water on the marine bench and thus permits waves once more to erode the base of the cliff. Marine planation would only be possible, therefore, on a subsiding land area. This view is expressed by von Richthofen[47] in his great work on China, where he states that slow depression alone can produce regional abrasion, since without progressive sinking the waves soon become exhausted on a narrow platform of their own carving. The same idea is expressed in his "Führer für Forschungsreisende"[48]. De Martonne, in his "Traité de Géographie Physique"[49]; de Lapparent in his "Traité de Géologie"[50] and his "Leçons de Géographie Physique"[51]; Kayser in his "Lehrbuch der Geologie"[52]; and Scott in his "Introduction to Geology"[53] are among the text-book writers who have adopted von Richthofen's theory that waves cannot cut far into the land unless wave erosion is aided by coastal subsidence. Many others have taken the same position, and some have even gone so far as to cite wave erosion as an indication of land sinking. For example, Hahn[54] says we must suspect a sinking of every region which suffers loss through the washing away of its margin, and Haage[55] gives wave erosion as one of the distinguishing characteristics of a sinking coast.

On the other hand, there are a few who have maintained that waves will continue to cut into any land mass so long as it projects above sealevel, whether or not it is undergoing depression. Ramsay, the first to recognize the power of the waves to produce a plane of abrasion, clearly expresses his belief that while subsidence of a land mass will aid the process of marine erosion, it is not essential; since, "taking *unlimited* time into account," any land area must eventually be worn away by the waves[56].

Green seems equally convinced of the ability of wave erosion to produce an extensive plane of denudation without subsidence, as he explains the origin of such planes without mentioning changes of level[57]. Both of these authors failed to appreciate the considerable depths to which wave action extends, Ramsay assuming that " the line of denudation " is "a level corresponding to the average height of the sea," while according to Green marine denudation must reduce a country to " an even surface coinciding approximately with the level of the lowest tides." Jukes-Browne[58] was almost as conservative in his estimate of the depth of marine erosion. Davis[59], Gulliver[60], and Fenneman[61] are among those who recognize not only the possibility of indefinite wave erosion on a stable land mass, but the additional fact that the baselevel of wave erosion is located at an appreciable depth below the water surface.

In the opinion of the writer any careful analysis of the process of marine erosion must lead to the conclusion that marine planation is possible without coastal subsidence. We have already seen that where the resultant of wave action is landward, material is driven toward the shore until the steepening of the shore profile produces a condition of equilibrium in which material driven up the slope by the landward-acting forces returns again under the combined influence of gravity, undertow, and other seaward-acting forces. During the entire period of equilibrium, sand, pebbles, and shingle are driven back and forth, up and down the beach slope, continually grinding themselves finer and finer. In this gigantic mill which borders the lands, rock fragments are continually reduced to a state of such exceedingly fine comminution that they are readily removed from the shore and shoreface zones as suspended particles in the water. During and after heavy wave action the water is turbid with matter in suspension to a considerable distance from the land. Part of this suspended matter is removed far from shore by the many currents which are involved in oceanic circulation, and finds a permanent resting place beneath the quiet waters of abysmal depths.

If there is an absolute loss of land where the resultant of wave action is landward, and the shore profile is built forward until equilibrium is established, how much greater must be that loss when the seaward components of wave agitation prevail and

PLATE XXVIII.

Marine cliff and wave-cut rock bench on the Pacific coast south of Cape Flattery, Washington. Note the very scanty beach deposit at the inner edge of the rock bench. Compare Plate XXII.

coarser débris on the bottom, as well as material in suspension, is transported seaward to be deposited over the edge of the continental shelf. Account must also be taken of the fact that agitation of the marine veneer is continually grinding its particles smaller and grinding material from the solid surface of the abrasion platform, thus producing fine sediment which currents may readily carry to deeper water during and after vigorous wave action. The never-ending shifting of the beach deposit back and forth over the shore and shoreface, already fully described, is accompanied by a ceaseless loss of the finest attrition products. Whether the shore profile is in equilibrium or not, whether waves are depositing beach material or sapping cliff bases, whether the marine veneer is increasing or decreasing in volume, there is a constant loss of very fine matter which is borne far away to deep water by current action. This means an eventual loss of equilibrium which must ultimately be restored by the erosion of more material from the lands. Where-ever there is wave erosion there is an absolute loss of material from the lands attacked. The laws of wave action afford no basis for Mitchell's conclusion that the sea restores to the continent " all the material washed from its bluffs and headlands "[62]. On the contrary, we must conclude that the sea never restores anything to the continent, except temporarily. The surface extent of lands temporarily built by marine agencies may be great, but their total volume above sealevel is small as compared with the volume removed by marine erosion. There was much of truth in the statement made nearly a century ago by Robert Stevenson, to the effect that " these apparent acquisitions are no more to be compared with the waste alluded to, than the drop is to the water of the bucket [63]". In the end the temporarily restored materials must themselves suffer removal by the combined action of waves and currents, which, however slowly, yet unceasingly destroy any land mass exposed to their attack. Under the most unfavorable conditions the loss from the lands will be small, but real. Where conditions favor vigorous wave erosion, rapid disintegration of the rock fragments, extensive solution of the rock-forming minerals, efficient transportation of the mechanical débris offshore, and high current velocities continued to deep water, the wasting of the land may be exceedingly rapid.

It is not possible that waves should exhaust themselves upon a platform of their own carving, and thus fail after a time to continue cliff erosion; for the loss of wave energy means that that energy has been expended upon the platform in question, and energy so expended can have but one result: abrasion and consequent lowering of the platform. This partially removes the cause of wave exhaustion, so that later waves reach the cliff base with enough energy remaining to effect some slight erosion. Any surface shallow enough to retard wave attack must suffer denudation until the attack is resumed. The vertical limit of marine denudation is a surface so low that wave action is no longer retarded by it. The corollary of this is that there is no horizontal limit of marine erosion.

In this connection it should be pointed out that fairly rapid wave cutting may occur at the base of a cliff which has been pushed far into the land. This arises from the fact that the shore profile must change extensively at times because of large variations in the forces attacking the shore. Imagine, for example, that on a shore which had been in nearly perfect equilibrium under gradually weakening wave attack for so long a time that the beach deposit had wasted away to a very small volume, a series of unusual storms should drive in vigorous oscillatory waves and develop a strong undertow. It is quite conceivable that the cliff, which had for years scarcely been touched by the waves, might be steepened and driven inland with comparative rapidity. In this case the *average* rate of cliff retreat would be exceedingly small; but the absolute rate for a limited time might be high. Variations in the direction and strength of longshore currents might also be accompanied by increased rate of cliff recession, in any place where such variations materially affected the condition of the shore profile. Even fairly rapid cliff retreat on a late mature or old shore profile is not, therefore, necessarily proof that coastal subsidence has admitted larger waves by deepening the water offshore.

Effect of Deposition.—One might suppose that the deposition of organic or chemical sediments from the water above the abrasion platform would protect the platform from erosion and by preventing its further deepening eventually stop further cliff erosion. But a little consideration will show that such deposition

PLATE XXIX.

Photo by S. W. Cushing.

Marine peneplane and monadnock near Madura, east coast of India.

can only bring about an elevation of the general level at which marine planation will occur. In the early part of the shore cycle, when deposition is small and erosion vigorous, the platform will be rapidly lowered. Later, when the increased depth of water over the platform permits more extensive deposition of organic or chemical sediment from the increased volume of superjacent water, and at the same time results in decreased intensity of wave erosion on the platform, the contending forces will be more nearly balanced. Equilibrium will be established and the effective wave base reached, when wave agitation and current action combined can just effect the removal of the deposits which tend to accumulate on the platform. Were it not for the burden of removing these deposits, the waves would reduce the platform still lower.

It is sometimes stated that before waves can erode cliffs far into the lands, rivers will bring out vast quantities of sediment which will in turn be widely distributed along the shores by currents. Deposition of the sediment, it is argued, will shallow the water, protect the shores and sea-bottom, and effectively prevent further cliff retreat. This argument assumes that the quantity of débris brought out by rivers and distributed along the margins of the lands is equal to or greater than the quantity of débris which the marine forces are competent to remove, and that therefore the entire energy of those forces is consumed in handling river-brought material. While it seems to the writer that as a general proposition this assumption is untenable, we may temporarily grant its reasonableness for sake of argument, providing it refers to a youthfully or maturely dissected land mass. As the land wears lower and the streams become more sluggish, the latter will bring to the sea a decreasing amount of sediment, an increasing proportion of which will be carried in suspension and so will be borne out to deep water without pausing in the vicinity of the shores. The forces of marine erosion and transportation will eventually remove the deposits which impeded wave attack during an earlier part of the fluvial cycle of land dissection, and once more the relentless encroachment of the sea will be manifest. Under the assumption least favorable to wave erosion, therefore, the progress of marine planation cannot be stopped by river-brought sediment. It can only be delayed. Deltas may be built seaward against the

PLATE XXX.

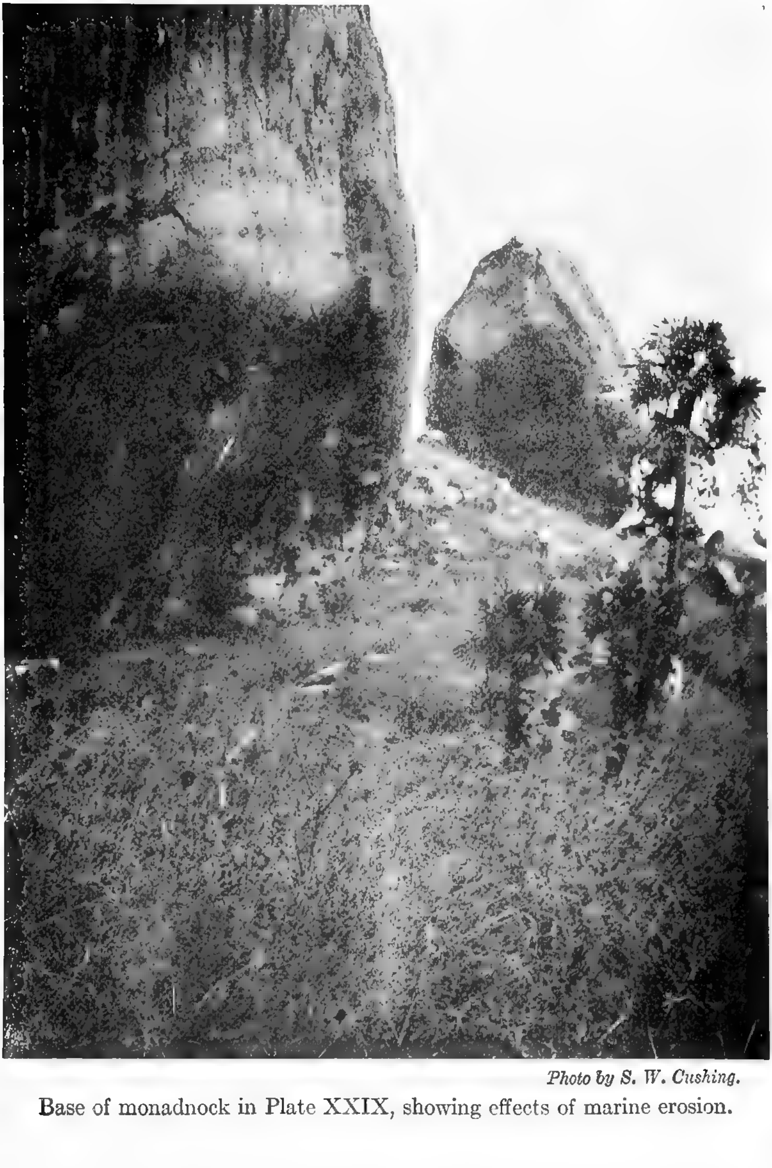


Photo by S. W. Cushing.

Base of monadnock in Plate XXIX, showing effects of marine erosion.

weves for a time, and help to keep parts of the shoreline young; late-mature coasts are delta-free.

The direct effects of river sediments in preventing cliff erosion have been exaggerated, as intimated in the foregoing paragraph. Near the mouth of many rivers it is perfectly apparent that the river deposits are directly shielding the cliffs from wave attack. But active cliffing is going on along many other coasts in spite of the fact that numerous streams enter the sea through valleys opening in the face of the cliffs; while long stretches of coast have enormous accumulations of beach deposits demonstrably not of fluvial origin. When we come to consider the deposits in the offshore zone, however, it is probable that greater importance must attach to stream-brought sediments, and that indirectly they may play an important rôle in certain stages of the marine cycle. Let us analyze, if possible, the relation of the marine cycle to the fluvial cycle of land dissection.

Correlation of the Marine and Fluvial Cycles. — Imagine a newly uplifted land mass of great areal extent and irregular surface, attacked at once by wave and stream erosion. During the youthful stage of stream development sediment is being eroded from some parts of the stream profile only to be deposited elsewhere as filling for lake basins, as alluvial fans, flood plains, and other temporary accumulations. Only a small part of the sediment reaches the sea. During this period the shore profile is in the young and perhaps early mature stages of its development, the abrasion platform is being rapidly developed, and the cliffs are being pushed steadily inland. Since river-brought sediment is small in amount at this time, the development of the shore profile is not greatly affected by it.

When the drainage system on the land is thoroughly integrated, all its parts nicely adjusted, and the stream profiles of equilibrium perfected, sediment from all parts of the land surface is ceaselessly swept seaward. At the river mouths the coarser sediment may be deposited in a delta, or driven along the shores in either direction. The finer material will be transported to a greater or less distance by some of the many types of marine currents, and much of it deposited in the offshore zone. By this time the marine cliff has been pushed well into the land, the abrasion platform has attained a considerable width, and a thin layer of marine veneer is journeying slowly down the sub-

marine slope toward the edge of the continental terrace. Increasing volumes of river-brought sediment are now deposited upon the shelf, adding to the thickness of the marine veneer and continental terrace, and thereby shallowing the water of the offshore zone. Aggrading will continue until a level is reached where the increased wave and current agitation is sufficient to remove the amount of débris which is deposited. The new profile of equilibrium will hardly rise to the surface of the sea, except under exceptional conditions in limited areas, as where deltas are temporarily formed. Coasts bordering shallow inland seas, or otherwise protected from the full attack of destructive marine forces, may, like the coasts of Holland and Belgium, be built far forward by delta accumulations before the inevitable period of their removal begins. Especially will this be the case if lands raised to mountainous heights shed débris into the sea through many rivers with exceptional rapidity. Both modern and ancient examples of coasts where conditions favored extensive delta growth are cited by Barrell[64], who fully recognized the temporary nature of the delta protection of coasts. Around most of the land margin, where delta protection is lacking, smaller waves will continue to traverse the waters shallowed by the deposition of river-brought débris, and will continue to erode the cliff, but more feebly than before. Cliff retreat, already slow because of the increasing breadth of the offshore zone, will be still further retarded by virtue of the decreased depth of water in that zone. Maturity of land drainage, therefore, means retarded shoreline development.

As the rivers of the land approach old age, they become more sluggish. Meandering in circuitous courses on a very low gradient, they can transport but a limited volume of the finest sediment. Decreased land relief is accompanied by decreased rainfall and increased loss of water by evaporation; and this means diminished stream volume. A smaller quantity of finer débris is weathered from the very gentle slopes of the old valley sides, to be carried to the sea by shrunken and enfeebled rivers. More material is removed in solution; less and finer material is removed in mechanical suspension. Upon reaching the sea much or all of this very fine material may be transported far from the land by marine currents before deposition is possible. Waves and currents in the offshore zone are no longer over-burdened

with river deposits, and expend their excess energy in removing the material previously deposited. The depth of the water is increased until the abrasion platform is again exposed to the slow wear of the migrating marine veneer. Larger waves gain access to the land, and cliff erosion is relatively more effective than before. But increased breadth of the abrasion platform and continental terrace compels material eroded from the cliffs to make a longer journey to deep water; and the longer time necessary to dispose of cliff débris necessarily tends to retard cliff recession. On the other hand, the reduction in land relief accomplished by the subaërial forces gives a lower marine cliff from which a smaller amount of débris is offered to the waves and currents for removal. Current action along the smoothed-out contours of mature and old shores may be much more effective than on the more irregular shores of youth; and the consequent more effective removal of débris from the shores may compensate, in part at least, for the greater distance to which it must be removed. All things considered, it seems to the writer that the retrograding of the shoreline must proceed more rapidly during the old age of land dissection than during its maturity. Especially must this be true where the reduction of extensive land areas by subaërial denudation causes a progressive rise of sealevel due to the infilling of sediments in the ocean basins. The significance of these relationships in the cycle of marine sedimentation has already been ably discussed by Barrell[65] in his essay on the "Relative Geological Importance of Continental, Littoral, and Marine Sedimentation."

The foregoing considerations lead to the interesting conclusion that, other things being equal, marine erosion should proceed most effectively about the low-lying desert areas of tropical regions, especially on the windward sides of such areas. For the absence of rivers would permit the development of the shore cycle, unretarded by any land sediments except the very fine material borne seaward by the winds. On the windward side even the æolian deposits would be lacking, while the onshore winds would continually drive vigorous waves against the cliffs, and by elevating the water surface would tend to produce a strong undertow which would assist in removing the products of wave erosion to deep water. On the leeward side the interfering action of wind-borne material, the prevalence of mild wave action

PLATE XXXI.

Photo by S. W. Cushing.

Monadnock on marine peneplane of the east coast of India.

because of offshore winds, and the existence of a landward instead of a seaward bottom current, would all tend to retard cliff recession. The absence of great storm waves in low latitudes, and the presence of coral building polyps, constitute special factors which would have to be taken into consideration in any attempt to compare shore development about tropical deserts with that about the more humid land areas of higher latitudes.

Independence of Marine and Fluvial Cycles. — It is important to remember that there is no necessary connection between the stages of development of a shoreline and the stages of development of the land mass which it borders. Each one develops independently, the one under the influence of marine forces, the other under the influence of subaërial forces. If both begin their evolution at the same time, the shoreline may be young while the land mass is in a youthful stage of development; and it may even happen that both attain full maturity at about the same time. But this is not a necessary, and not even a common relation. When a young shoreline of submergence is produced by the partial submergence of a mature land mass, the land mass remains mature throughout the youth of the shoreline, for a slight submergence, which is sufficient to initiate an entirely new cycle of shoreline development, produces scarcely any appreciable effect upon the main mass of the land. The sea invades the lower reaches of the valleys, and the remaining lower courses of some rivers may have their gradients slightly reduced if delta building takes place at the bay heads. But the land mass as a whole still consists of high hills and ridges separated by deep-cut branching streams; it is still a maturely dissected region, and its cycle of erosion continues without any real interruption toward the ultimate goal of planation.

The principle here involved is an important one, and since there is not complete agreement concerning it, a further word of explanation is in order. Davis has at different times presented the idea that any change of level introduces a new cycle of land-mass development. According to his interpretation the land mass which was mature before depression had inaugurated a new cycle of shoreline development, would become young in a new cycle of subaërial erosion as soon as the change of level occurred. In speaking of such changes of level he writes, " The previous cycle (of land dissection) is thus cut short and a new cycle is

entered upon"[66]; and again, "a cycle is interrupted when the land mass rises or sinks, or when it is warped, twisted, or broken. Like accidents, interruptions may happen at any stage of development. It is then convenient to say that the sequential form attained in the first incomplete cycle shall be called the initial form of the new cycle, into which the region enters, more or less tilted or deformed from its former shape"[67].

There are certain theoretical considerations which favor such an interpretation as is outlined in the above quotations; but it seems to the writer that numerous practical difficulties outweigh these considerations. Under the proposed scheme a submature plateau with large, flat-topped inter-stream areas, a mature plateau with sharp-crested ridges separated by V-shaped valleys, and an old plateau characterized by low and gently undulating topography, would all have to be called "young" in case each had been slightly depressed and not much modified since. Forms of totally different appearance, and typically characteristic of three distinct stages of normal plateau dissection, would be grouped together as in the same stage of development in the new cycle due to submergence. An observer in the interior would never be able to tell the stage of development of land forms until he had visited the coast to make sure that neither emergence nor submergence had introduced a new cycle, the recognizable effects of which were limited to the coastal zone. Indeed, he would find that practically all land masses are young in the current cycle, for emergence or submergence has occurred on most coasts within a period geologically so recent that little modification of surface forms has occurred since. The terms young, mature, and old would no longer be aids to an appreciation of significant differences in land forms, and the strongest argument for interpreting the surface features of the earth in terms of their stages of development would disappear. Davis has himself in a recent volume[68] recognized the difficulty of applying strictly his earlier suggestions regarding the terminology of the cycle and has proposed to avoid the difficulty in part by the use of circumlocutions or explanatory paraphrases.

If it appears that I have pushed an unimportant point to an absurd extréme, it must be remembered that a substantial agreement as to the usage of the terms cycle, young, mature, and old is absolutely essential to an intelligent understanding of

land form description, and that the difficulties I have portrayed are the necessary and logical consequence of considering every change of level as inaugurating a new cycle of land-mass development. If the same mountain mass is to be called mature by one observer because of the advanced stage of its dissection by stream erosion, and young by another observer who finds that its borders are slightly submerged in the sea, endless confusion must result. It will scarcely meet the situation to say that "a mature mountainous region was slightly submerged and is now young in the new cycle," for such a double description is too cumbrous to supply the need for a concise, clear, and consistent method of land-form description. One may, however, properly say that "a mature mountainous region was slightly submerged, and its *shoreline* is now young." Every significant change of level does introduce a new cycle of shoreline development; and it is evident that failure to attach sufficient importance to the fact that the cycle of shoreline development and the cycle of land-mass development are wholly independent, and progress at different rates under the influence of different forces, is responsible for the conception that every change of level introduces a new cycle of subaërial denudation. The absence of a clear discrimination between the two cycles is especially noticeable in the context from which the second of the above quotations is taken[69].

All of the difficulties discussed above disappear if we adopt the following as fundamental principles in land-form description: (1) The cycle of shoreline development and the cycle of land-form development are measurably independent as regards the evolution of their sequential stages, and must be treated as two distinct cycles. Both may originate from the same change of level, their corresponding stages of development may in some instances be closely correlated especially near the sea, their relative rates of progress in a given region may be compared, and the influence of one upon the other may be studied; but the two distinct cycles must always be carefully discriminated. It might be added that the cycle of stream development, and the cycle of land-mass development (dissection) which are very generally confused with each other, as well as with the marine cycle, are likewise distinct and may progress at different rates[70]. (2) Emergence introduces a new cycle of shoreline development, and will, if of sufficient magnitude, introduce new cycles

of stream development and of land-mass dissection. A slight emergence, especially if very gradual, will merely accelerate the progressive development of the cycles of stream development and of land-mass dissection already current, or will, if repeated, cause pulsations of reinvigorated stream cutting to advance inland up the rivers. The result may be minor topographic changes of the highest importance to the student of past fluctuations of level, and these topographic records must be fully appreciated and emphasized. But unless they are of large magnitude, rising to the dignity of a truly rejuvenated topography, the short episodes which they represent should not be dignified by the name of cycles. (3) Submergence introduces a new cycle of shoreline development, but submergence alone never introduces a new cycle of stream development or of landform dissection. This is because the forces which cause stream development and land-mass dissection continue their work as before, in essentially the same relative positions as before, even though absolute altitude is different and absolute efficiency may be more or less modified by a change in rainfall. Bayhead deltas may form in the drowned valleys, the gradients of some streams may be diminished for a limited distance inland from their mouths, and the rate of erosion on adjacent slopes may be somewhat retarded. But these local and temporary effects have no appreciable influence on the dissection of the land as a whole, and to no extent do they "rejuvenate" it. In other words, submergence does not "determine a more or less complete break in processes previously in operation, by beginning a new series of processes with respect to the new baselevel"[71], and therefore does not inaugurate new cycles of stream or land-mass development.

It is important that one should distinguish, not only between the shoreline cycle and other physiographic cycles, but also between stages in the development of the *shore profile* and stages of *shoreline* development. On a shoreline of submergence, for example, it very often happens that the shore profile at certain points becomes mature (*i.e.*, the marine bench is pushed inland, the cliff weathers back to a gently inclined, soil-covered slope, and the shore profile of equilibrium is fully established) long before the shoreline as a whole is reduced to a comparatively simple line back of the bayheads. The shoreline in this case is still young, and may even retain the excessive irregularities of

very early youth; but the shore profile is mature at some places, although still young at others. On the other hand, where wave erosion is unusually effective the irregularities of a shoreline of submergence may be quickly removed, and the shore outline transformed to a line of simple curvature back of the original positions of the bayheads, at a time when the waves are still actively undermining the marine cliffs and pushing them backward. In this case the shoreline is mature, but the shore profile is young. It would be quite proper to speak of that part of the profile called the marine cliff as a "young cliff," but to describe the shoreline as "young" would be erroneous.

The coast of Normandy exhibits a shoreline of fairly simple curvature, bordered by very steep or even vertical cliffs of bare rock, from which landslides often descend into the rapidly advancing sea. Here the shoreline is mature or late mature while the profile is young. Davis[72] has described the cliffs of this coast as late mature ("spätreife Kliffe"); but it is evident from his descriptions and from his characteristically expressive diagram representing this coast, that the expression "late mature" really applies to the shoreline alone, while the cliffs along the shoreline are marked by the steep slopes and frequent landslides found only in young cliffs. Davis has himself given such excellent accounts of the diverse features of young and late mature marine cliffs in other connections that there can be no doubt his application of the term "late mature" to the Normandy cliffs was merely an oversight such as is common in physiographic literature where stages of shoreline development and stages of shore profile development are not sharply discriminated.

Comparative Rapidity of Marine and Fluvial Planation. — Among those who admit the ability of unlimited wave action to reduce a land mass to an abrasion platform below sealevel, there are a number who believe that fluvial denudation takes place so much more rapidly, that any large land mass must be reduced to a peneplane before the waves could cut any great distance into the lands. This view is well expressed by Geikie[73] in these words: "Before the sea, advancing at the rate of ten feet in a century, could pare off more than a mere marginal strip of land, between 70 and 80 miles in breadth, the whole land might be washed into the ocean by atmospheric (meaning fluvial) denudation."

It is admittedly a difficult matter to find any basis for an adequate comparison of the relative rates of marine and fluvial denudation; but there should be no difficulty in seeing that Geikie's comparison is based on figures which enormously overestimate the average rate of stream erosion. As a starting point in his calculations he takes the amount of sediment annually discharged by the Mississippi River, and computes that this river will lower the land throughout its whole drainage basin an average of 1 foot in 6000 years. He cites other rivers which reduce their drainage basins much more rapidly, but the rate just given is assumed as a conservative figure in calculating rates of fluvial denudation. The error consists in reckoning denudation at the same rate throughout the entire fluvial cycle. It is true Geikie recognizes that "the last stages in the demolition of a continent must be enormously slower than during earlier periods"[74], but he makes no allowance for this fact in his calculations, except to intimate that the resulting error may be compensated for by the material removed in solution and not figured in the above estimate.

The Mississippi River drains vast areas of high mountains and plateaus whose steep slopes contribute large quantities of waste to its upper branches; and extensive stretches of semi-arid plains where fine-grained unconsolidated sediment is shed into the streams with enormous rapidity. Much of the river's drainage area has reached maturity, and its larger branches are transporting heavy loads of débris on fairly steep profiles of equilibrium. There can be no comparison between the amount of material carried to the sea by the Mississippi at the present time, and the amount which will be carried when the mountains and plateaus are worn lower, the stream gradients reduced, the rainfall diminished because of decreasing relief, and the stream volumes greatly lessened because of decreased rainfall and increased evaporation. The annual denudation under those conditions will be but a very small fraction of what it is to-day, unless the efficiency of æolian denudation is enormously increased as the land wears lower. Instead of allowing 4,500,000 years for the removal of the entire continent of North America, it is conceivable that it might be nearer the truth to allow that much time for the reduction of the surface by 1 meter during the latest stages of subaërial denudation. Portions of tertiary pene-

PLATE XXXII.

Imposing marine cliff at North Cape, Norway

planes which have been exposed to erosion for a period which may be estimated as one or more millions of years[75], not only have not been reduced nearly to sealevel, but seem to stand somewhere near the original positions of the upland surfaces. Very many more millions of years would be required to reduce these areas of hard rock to a low-lying surface of fluvial denudation. How much this time might be shortened by æolian erosion is problematical; but the combined action of the subaërial forces could scarcely accomplish the work in so short a time as a few million years.

It appears, therefore, that while it is not possible to more than guess at the time required for the subaërial denudation of a continent, the advantages are not so overwhelmingly in favor of subaërial denudation, and against marine denudation, as has been supposed to be the case. There are indeed, as we have already seen, certain marked advantages in favor of marine planation, not the least of which is the slight rise of sealevel, due to the infilling of sediment in the ocean basins and therefore normal to the marine cycle, which brings the waves against the non-resistant fluvial deposits and residual hills of the old land mass. So far as *a priori* reasoning is concerned, we should recognize the possibility that wave erosion may completely plane away a large land area before the subaërial forces have had time to reduce it to sealevel. Which forces have been the more effective in producing known peneplanes must be decided, if at all, on the characteristics of the peneplanes themselves, and not on the basis of *a priori* arguments.

Probability of Marine Planation. — Several authors have expressed the opinion that movements of a land mass must prevent extensive marine planation by repeatedly forcing the waves to begin anew the cycle of denudation at a new level. This view is stated by Davis in the words: "The sensitiveness of a local shoreline to changes in the ocean basin or border all around the world makes extensive plains of marine abrasion of improbable occurrence"[76]. Emphasis is properly laid upon the fact that marine erosion is restricted to a narrow vertical zone about the margin of the lands, the position of this zone altering with every change in the relative level of land or sea; whereas subaërial denudation goes on simultaneously over the entire land area, regardless of sealevel oscillations. "A slight movement of

elevation usually sets the sea back to begin its work anew on the seaward side of its previous shoreline, but such an elevation only accelerates the work of subaërial denudation all over the elevated region. The waves on the seashore shift their line of attack with every slight vertical movement of the coastal region; but the subaërial forces over large continental areas gain no notice of slight movements until a considerable time after they have been accomplished, and hence they perform their task only with reference to the average attitude of the land"[77].

When the theory of fluvial peneplanation was first proposed, it was objected that the land could not stand still long enough to permit streams to wear large areas nearly down to sealevel. The best answer to this objection was the finding of broad erosion surfaces, the characteristics of which indicated a fluvial origin. In like manner, we must depend upon field evidence to settle the question whether extensive planes or peneplanes of marine denudation have been produced in the past. We may fully recognize the sensitiveness of marine erosion to changes of level, without denying the possibility of marine planation. If we find erosion planes having the characteristics of planes of marine abrasion rather than those of subaërial denudation, we may reasonably conclude that the land can stand still long enough for waves to reduce a land mass to a plane surface. The distinguishing features of planes and peneplanes of different origins have been receiving more attention in recent years than formerly, and we may anticipate that discrimination between these surfaces, at least where they are fairly well preserved, will become practicable.

Attention may here be called to a tendency to regard erosion surfaces which show characteristics of marine planation, as fluvial peneplanes which have been planed down further by the sea. It would perhaps be more pertinent to speak of them as marine planes or peneplanes whose development was favored by extensive subaërial erosion of the land. For when we remember that relatively flat fluvial peneplanes may have a relief of several hundred feet and that marine abrasion reduces a land mass many feet below sealevel, it is evident that the waves must perform much work in planing away a fluvial peneplane. Furthermore, marine abrasion destroys the essential characteristics of fluvial denudation, including the extensive adjustment of stream val-

Rocky headland and drowned valley of a young shoreline of submergence, near Clifton, Massachusetts

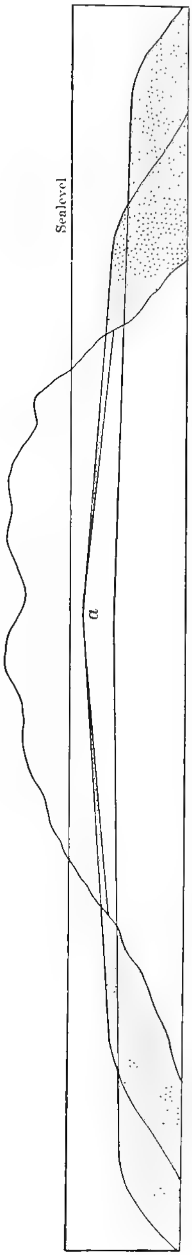


Fig. 40. — Stages in the reduction of a land mass, first to a flat cone covered with marine veneer, later to a still flatter rock cone free from veneer but bordered by a broader continental terrace.

leys to weak rock belts, which is one of the best evidences of long-continued fluvial action. Since both marine and fluvial planation are possible, it is perhaps safer, in the absence of evidence to the contrary, to regard an erosion surface covered with remnants of a marine veneer as a marine plane or peneplane, rather than to make the gratuitous assumption that there must have been a fluvial peneplane which was later planed off by the waves. Rapid depression of a fluvial peneplane would economize the amount of wave work necessary to produce the observed result[78]; but one is not justified in assuming both fluvial peneplanation and rapid submergence in the absence of supporting evidence.

It is sometimes assumed that a cover of marine sediments is an essential feature of a marine plane or peneplane[79]. While a thin marine veneer may be expected in many or even in most cases, its presence in any appreciable quantity does not seem necessary in the later stages of the marine cycle. A land mass reduced to an abrasion platform surrounded by a continental terrace, as a result of wave attack from all sides, would, in the penultimate stage, have the form of a very flat cone with the apex (*a*) where the last land surface was reduced (Fig. 40). Unimpeded by any further contributions of débris from a land area, wave erosion would proceed to remove the veneer which might have accumulated on the faintly conical platform, and to reduce the rock surface to wave base with no cover except an insignificant amount of recently eroded débris in transit to deeper water. As explained on a previous page, the conti-

nental shelf would then consist of deposited material at the outer borders, and a bare rock erosion surface within. Essentially the same conditions might prevail at an earlier stage, where a broad continental shelf bordered a still remaining land area, if the supply of land waste were very slow. When uplifted the land area, abrasion platform and continental terrace would occupy the same relative positions as the Older Appalachian Mountains, the Piedmont Belt, and the Atlantic Coastal Plain. A very thin marine veneer, quickly removed, might be insufficient to superimpose rivers upon transverse hard rock ridges, even when these were worn down nearly to the level of an almost plane abrasion surface; and this partial initial adjustment of streams to rock structure would be greatly increased during further dissection. It is essential, therefore, to keep an open mind as to the possible origin of uplifted and dissected peneplanes which show no traces of a former marine cover, and which may even show a considerable adjustment of stream courses to rock structure.

Interruptions and Accidents During the Marine Cycle. — Davis[80] has repeatedly emphasized the importance of the "interruptions" and "accidents" which frequently occur in the fluvial cycle. Similar events diversify the history of the marine cycle. We have already seen that elevation may end the progress of the marine cycle at a given level by raising the abrasion platform and continental terrace, or parts of them, above the reach of the waves. Subsidence, if rapid, may produce the same effect by lowering the platform and terrace far beneath the lowest limits of wave activity. Slow, progressive subsidence may simply hasten the development of the marine cycle by constantly deepening the water offshore and thus facilitating wave erosion. It would seem, however, that any considerable help from subsidence would demand rather rapid sinking, in order to keep the water offshore continually and appreciably deeper. As subsidence progresses the inner margin of the continental terrace advances landward, so that the outer margin of the abrasion platform is progressively overlapped by a wedge of marine deposits which thicken seaward (Fig. 41).

Accidents may occur during any part of the marine cycle, and locally interfere for a time with the normal development of the shore profile. Glaciers may excavate deep troughs far below

wave base. Volcanic eruptions may build cones upon the continental shelf, the summits of the cones possibly rising above sealevel. But in course of time the submarine troughs will be filled with sediment, the volcanoes will be removed by wave erosion, and the development of the shore profile will proceed

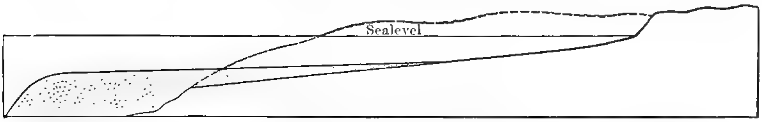


FIG. 41. — Overlapping of marine deposits upon the abrasion platform of a slowly subsiding land mass.

as before. A longer-enduring departure from the ideal scheme will occur if a strong and deep ocean current abrades the bottom long enough to reduce it below wave base. But even this accident must be corrected as the removal of land masses and the reduction of shallows to wave base make concentrated current action impossible.

SHORELINES OF EMERGENCE

Initial Stage. — In the typical shoreline of emergence the water margin comes to rest against the exposed sea floor. Under normal conditions this floor consists of an abrasion platform and continental terrace, the smooth surface of which is intersected by the plane of the sea surface to form a very simple shoreline. Inland the land rises very gently in the form of a

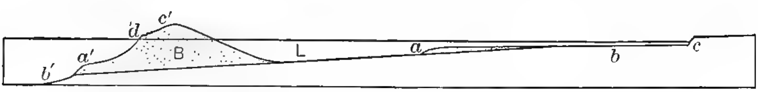


FIG. 42. — Elements of the profile of a shoreline of emergence.

smooth marine plane or coastal plain, as the case may be; seaward the bottom slopes downward with the same gentle inclination, giving shallow water for a long distance offshore.

Davis has briefly outlined the broader features in the development of shorelines of emergence on the assumption that emergence is relatively rapid and is then followed by a still-stand of the land. Waves attack the initial shoreline with results to the profile which are at first similar to those produced at the shoreline of submergence. A marine bench (Fig. 42, *b*) is cut, a marine

cliff (*c*) produced, and a shoreface terrace (*a*) built. The bench may be covered by a thin beach deposit. But in two respects the development of the profile of a shoreline of emergence is significantly different from that previously described. In the first place, only small waves can reach the shore, because, according to the law of wave breaking set forth in Chapter I, large waves break when they enter water whose depth is approximately equivalent to the wave height; and this must occur well out from land in the case of shorelines of emergence. Accordingly, the marine bench is shallow, and the marine cliff is low, both being pushed slowly into a low lying plain by weak waves. Because of its insignificant size, the cut made by the waves during this earliest stage of development is often spoken of as a *nip* in the edge of the land. The nip is frequently preserved from further change for a long period of time by the development of an *offshore bar* (*B*), which is a second feature characteristic of the shoreline of emergence, not found along typical shorelines of submergence.

As is elsewhere pointed out, it may well happen that progressive emergence prevents the formation of a distinct nip on the mainland shore until after the offshore bar has been formed. If the levels of land and water then become stationary, and the lagoon is sufficiently broad and deep, lagoon waves may produce a nip of later date than the bar. On the other hand, if emergence continues, or if submergence intervenes, or if the lagoon waves are too feeble, the nip may be entirely lacking. Whether or not a nip is formed, the shoreline is past its initial stage and entered upon the stage of youth as soon as the offshore bar is built up above the water surface.

Young Stage. — Under the various names of barrier beach, sand reef, and offshore barrier, the offshore bar has been described as a continuous narrow ridge of sand, lying some distance out from shore. Its seaward side has the normal beach profile of equilibrium, and its crest rises a few feet above high tide level.

The precise manner in which the offshore bar originates is not definitely known. Various theories advanced to account for its development are considered at length in a later chapter but only two deserve special mention here. The first is that of Gilbert, which is based on the belief that the material of the bar consists of "shore drift," which is being moved parallel to the

coast by longshore currents. "The most violent agitation of the water is along the line of breakers; and the shore drift, depending upon agitation for its transportation, follows the line of the breakers instead of the water margin. It is thus built into a continuous outlying ridge at some distance from the water's edge."[81] De Beaumont[82] would derive the material of the bar from the offshore deposits, by direct wave action. Davis, who follows de Beaumont, states the theory thus: "When waves roll in upon a shelving shore, much of their energy is expended on the bottom. Between the line of their first action far offshore and their final exhaustion on the coast, there must be somewhere a zone of maximum action. This zone must lie farther seaward when large storm waves roll in than when the sea is slightly ruffled in fair weather. . . . Here the bottom is deepened; the coarser particles are moved landward, forming a shoal and in time a bar inclosing a lagoon; while the finer particles are moved seaward, where they are distributed in moderate thickness over a considerable area."[83] Conformable to these two theories, Gilbert illustrates his idea of the offshore bar by a section which shows the bar deposit resting on the unbroken surface of an inclined sea-bottom; whereas in Davis's illustrations the sea-bottom is represented as deeply eroded by the waves which used the eroded materials to build the bar. In Gilbert's opinion the offshore bar is "absolutely dependent on shore drift for (its) existence. If the essential continuous supply of moving detritus is cut off, . . . the structure (is) demolished by the waves which formed it"[84]. According to Davis, offshore bars "might be developed essentially under the control of on- and offshore action alone"[85].

Without pausing to discuss the relative merits of these two theories at this time, we may note that the further development of the shore profile would be essentially the same in either case. The profile of the seaward side of the bar is a profile of equilibrium which varies with variations in the waves and other forces which affect the shore, in the manner already fully described. Beach materials are heaped upon the backshore (*d*) one day, and dragged out to form a shoreface terrace (*a′*) the next. Vigorous wave action cuts into the sea-bottom to form a marine bench (*b′*), while the top of the bar or the sand dunes upon its crest may have a low but distinct marine cliff (*c′*) marking the upper limit of the shore.

Normal development involves slow retrogression of the shoreline, as the grinding of the beach materials to fine silt permits their removal in suspension to deep water, or as seaward bottom currents drag coarser débris from the face of the bar down the inclined slope of the bottom toward the edge of the continental terrace. But the retrograding process does not necessarily involve the rapid removal of the bar. The material lost from the bar in the ways described above may be compensated for by material freshly cut from the sea-bottom during the landward cutting of the marine bench. Storm waves hurl débris over the crest of the bar to its back side, and the overwash of waves carries much additional material down its landward slope. Wind-blown sands still further assist this landward building. All these factors combined may be sufficient to build up the inner side of the bar as fast as the outer side is cut away, in which case the bar will retreat bodily toward the coast without any marked change in its average width.

Between the offshore bar and the mainland lies a narrow strip of shallow water, called the *lagoon* (*L*), whose weak waves faintly cliff the lagoon shores, often at a lower level than the initial nip. Tidal currents bring fine sediments from the surf-beaten outer side of the bar, to deposit them in the quiet water of the lagoon, which also receives some stream-brought sediment from the lands, wind-blown sands from the beaches and dunes of the bar, and débris eroded from the lagoon shores by the waves. In course of time these sediments may build the floor of the lagoon up to such a level that salt marsh vegetation can take possession in the manner described by Shaler[86] in his oft-quoted paper on the "Sea Coast Swamps of the Eastern United States," and so transform the lagoon into a salt marsh. It must not be supposed, however, that all salt marshes back of offshore bars have had the history outlined by Shaler; for, as will be shown in a later chapter, the typical salt marshes of the Atlantic Coast have been formed in an entirely different manner.

As the retrograding of the offshore bar continues, its sands and gravels are driven in over the marsh surface. The enormous weight of the bar compresses the peat and other marsh deposits, which later outcrop on the seaward side of the bar near or below low-tide level, and thus bear witness to the retrograde movement of the outer shoreline. During all this movement the

profile of equilibrium is maintained as perfectly as the varying conditions will permit. The bench is deepened as well as cut landward, and its seaward edge grades imperceptibly into a constantly broadening abrasion platform. Erosion products accumulate in a continental terrace farther seaward. At length the bar is driven upon the mainland, the marsh or lagoon is extinguished, and larger waves working on a steeper profile attack the coast where long before small waves on the gently sloping initial profile cut the less prominent nip. The shore profile is now thoroughly mature.

It is not necessary that the offshore bar should begin to retreat as soon as formed. Larger storm waves may build successive additional bars in deeper water on the seaward side of those formed earlier; but prograding of the shoreline from this cause can proceed to a very limited extent only, and the extensive series of "beach ridges" often attributed to this action must be explained in some other manner. One other explanation involves the supply of large volumes of débris by longshore currents, which will cause long-continued prograding in the manner already explained for shorelines of submergence. If the longshore currents supply just enough débris to make good the loss from wave erosion, attrition, and removal, the shoreline will remain stationary.

Mature and Old Stages. — Whether or not the offshore bar is prograded for a period, retrograding must inevitably replace the temporary forward movement in the course of time, and the shoreline be driven back upon the mainland. Maturity begins when the lagoon or marsh is extinguished, and the waves have begun their real attack upon the coast. From this time on there are no features of shore profile development which differ in any essential respect from the mature and old profiles on shores of submergence. As both these stages of profile development have been fully discussed in connection with shorelines of submergence, we may dismiss them without further consideration.

NEUTRAL SHORELINES

The successive stages of development in the profiles of neutral shorelines involve little that is novel save in matters of detail. Marine erosion of delta shorelines, alluvial fan shorelines, and

outwash plain shorelines would give stages resembling those in the profile of shorelines of emergence, except that the offshore bar stage need not necessarily be represented in case the seaward portion of the profile descends too abruptly into deep water.

The typical delta consists of two main portions, a subaërial plain and a subaqueous plain, separated by a steeper wave-cut slope to which Barrell[87] originally gave the name "shore face." The comparatively steep frontal slope of the delta may thus be far from the shoreline, as in the case of the Nile delta, and is unrelated to the true delta shore profile. The shoreface, on the other hand, is the steeper, landward portion of the shore profile of equilibrium, of which the profile of the gently sloping subaqueous plain is the seaward continuation. It should be noted that the outer margin of the subaqueous plain, where it joins the steeper frontal slope of the delta, does not mark the position of wave base, as most writers erroneously assume. It may mark the seaward end of the profile of equilibrium in any given section, the equilibrium referred to being the balance between the power of the waves on the one hand, and the work they must accomplish in transporting débris on the other. Stop the addition of sediment to the delta for a time, and the waves will slowly reduce the submarine plain, including its outer margin, to a still lower level. Where the surface of the water body in which a delta is built has recently been raised or lowered, the outer margin of the subaqueous delta plain is not only unrelated to wave base, but is also unrelated as yet to the normal profile of equilibrium for the new conditions. Wave base is an imaginary horizontal plane marking the lowest limit of effective wave erosion in a given water body. It is highly improbable that the seaward margin of any present day delta or shore terrace coincides with that imaginary plane, just as it is highly improbable that any present land surface coincides with the imaginary subaërial baselevel plane.

Neutral volcano shorelines would have the same profile development as slopes of corresponding steepness on shorelines of submergence. Coral reef shorelines have one striking peculiarity, in that they depend on organic as well as on inorganic forces for their history. Vigorous coral growth may indefinitely postpone the developmental stages of the reef under

marine erosion, and may even for a long period build the reef forward into the sea despite the most vigorous wave attack.

Fault shorelines deserve more than passing notice because of certain novel features which they present both in the initial and in later stages. If the hade of the fault plane is steep and the seaward block drops well below sealevel, in the initial stage the sea will come to rest against a steep cliff, the fault scarp (a^1a^2, Fig. 43) which descends abruptly into deep water. This initial stage may persist for an abnormally long period of time,

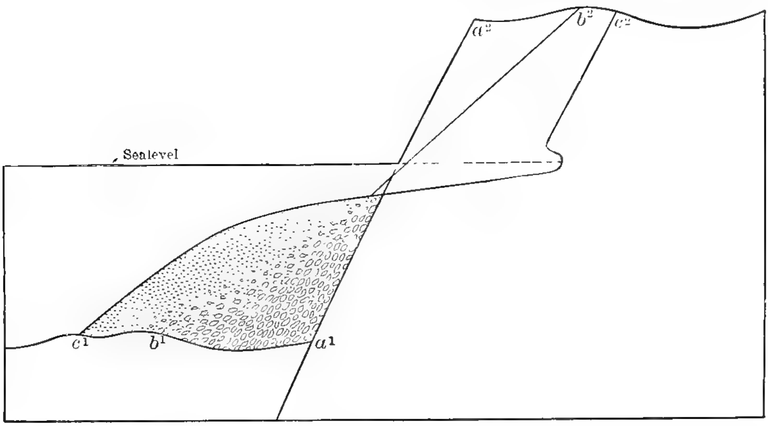


FIG. 43. — Stages in the development of the shore profile of a fault coast.

due to two important facts. In the first place, as we have already seen in an earlier chapter, waves approaching a vertical or nearly vertical wall rising out of deep water are reflected back without developing any great erosive power; and in the second place, where the water is deep close to shore the waves cannot arm themselves with any tools with which to facilitate their attack upon the land. Rock fragments weathering from the face of the cliff descend at once to deep water, beyond the reach of effective wave action. If the cliff is composed of very resistant rock which yields but slowly to the forces of weathering, the initial profile may long remain practically unaltered.

In the course of time, weathering of the cliff face causes it to retreat and leads to the accumulation, at its base, of a submarine talus (b^1). Two important consequences follow. Wave reflection is less perfect and hence the waves develop greater erosive

power on the more sloping surface of the talus; at the same time the waves become armed with the talus débris, which is hurled against the cliff face with ever-increasing force. Under these favorable conditions the retrograding of the cliff face may be so accelerated as to give it a steeper slope (c^2) than it possessed a short time before (b^2), while the prograding of a true shoreface terrace (c^1) replaces the former talus growth. From this time forth the shore profile develops as in the case of shorelines of submergence.

COMPOUND SHORELINES

The name " compound shoreline " has been applied to a shoreline which shows with more or less equal prominence features characteristic of at least two of the three simple classes of shorelines. The best examples of compound shorelines exhibit the irregular pattern of drowned valleys in combination with a smooth and gently sloping sea-bottom from which an offshore bar usually rises to the surface. There is reason to believe that in such cases extensive emergence takes place first, and that later moderate submergence drowns the valleys carved in the emerged coast. Were submergence to occur first, it is probable that partial emergence soon after would find the sea-bottom still possessed of its former irregularities to such a degree that the new shoreline would still be a typical shoreline of submergence, with its essential characteristics little affected by the uplift which operated merely to reduce the *amount* of submergence; whereas emergence a long time after would reveal a well-smoothed sea-bottom and give a typical shoreline of emergence. Compound shorelines, of which the North Carolina coast is a typical example, may therefore be regarded as presumptive evidence in favor of emergence followed by partial submergence.

The character of the initial profile of a compound shoreline of the North Carolina type will depend partly upon the amount of dissection which the emerged area experienced previous to the partial submergence, and partly upon where the profile is taken. If dissection was limited to areas adjacent to the main streams, and the profile is located so as to lie wholly in an undissected inter-stream area, it will not differ from the initial profile of the ordinary shoreline of emergence, providing no offshore bar has formed. When, however, an offshore bar forms before submergence changes the shoreline to the compound type, and this

bar is built up to the surface as submergence progresses, what may be called the initial profile of the compound shoreline will resemble the young profile of the shoreline of emergence, in which the bar is a prominent feature. If dissection was so extensive that submergence everywhere brings the water to rest against relatively steep valley sides, or if the profile is so located as to cross the shoreline of a slightly dissected and embayed plain within the limits of one of the drowned valleys, the initial profile of the compound shoreline will differ from the young profile of a normal shoreline of emergence in having a steeper slope at the water line, deeper water offshore, and a more irregular bottom for a limited distance seaward (Fig. 44).

Fig. 44. — Profile of a shoreline of emergence when sealevel is at *a*, changed to profile of a compound shoreline when submergence brings the sealevel to *b* and facilitates the landward migration of the offshore bar.

In case submergence is so rapid or so extensive as to destroy the original offshore bar, no new bar will form on the submerged irregular surface of the dissected land mass (unless the hills of the land were of such very moderate relief as to constitute practically a level plain), and we will have a normal shoreline of submergence instead of a compound shoreline.

During submergence the offshore bar may be driven landward by the larger waves which are admitted by the deepening water offshore. The small waves in the lagoon will faintly cliff the lagoon shores, and currents will proceed to smooth out the inequalities of the bottom by distributing the wave-eroded débris and the sediments brought in by tides and rivers. The further development of the shore profile will be similar to that of the ordinary shoreline of emergence.

On a compound shoreline combining the features of a fault shoreline with those of a shoreline of submergence a profile through one of the drowned valley sections will have in general the same developmental history as the normal profile of a shoreline of submergence. A profile through a typical portion of the fault scarp will pass through the sequential stages already described for normal fault shorelines.

RÉSUMÉ

We have now traced the history of the shore profile from its initial to its ultimate stage. The characteristics of the profile in all the different stages of development and in the several classes of shorelines have been fully considered, and shore profile development has been compared with the development of stream profiles. This study has led us to certain important conclusions, which must have an important bearing upon all investigations of marine erosion. Thus, it has been shown that a shore profile of equilibrium is early established, the maintenance of which is accompanied by constant loss of débris and consequent recession of the shoreline. The changes in this profile, which have given rise to so much misunderstanding on the part of many observers, are due to temporary changes in the balance of the shore forces, and are of small importance as compared with the general cycle of shore development. Of very great importance is the fact that long-continued wave action must reduce broad land areas to a plane, or at least to a peneplane, of marine denudation. We have found that such a plane or peneplane may be produced without progressive subsidence; that rapid wave cutting is no proof of a change in the relative level of land and sea; and that while subaërial denudation may temporarily embarrass marine abrasion by delivering much sediment to the sea margins, ultimate marine planation cannot thus be prevented. A comparison of the relative rapidity of marine and fluvial planation indicates that the widespread opinion in favor of the greater efficiency of fluvial erosion rests upon an inadequate basis, and that marine forces may really be able to reduce a large land mass to a peneplane more rapidly than can stream erosion. Whether the land stands still long enough for such a result to be effected by the waves, is a question which cannot be answered on *a priori* grounds and which depends upon careful and unprejudiced study of actual peneplanes for its solution. In prosecuting such study it is essential to remember that neither the absence of marine sediments nor the presence of a certain degree of stream adjustment is conclusive evidence in favor of a subaërial as opposed to a marine origin for a given peneplane. The marine cycle of erosion is subject to interruptions and accidents, the occurrence of which does not, however, affect the general principles controlling the cycle of shore development under normal conditions.

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CHAPTER VI ✓

DEVELOPMENT OF THE SHORELINE

A. SHORELINES OF SUBMERGENCE

Advance Summary. — It is the purpose of the present chapter to trace the systematic development of the shoreline of submergence from its initial stage of extreme irregularity and complexity until it acquires the regular and simple outline characteristic of full maturity. The features of old age are reserved for attention in a later chapter. Special consideration is given to those elements of shore form normally associated with the stages of youth and maturity, such as beaches, spits, bay bars, looped and flying bars, tombolos, cuspate bars and forelands, marsh bars, and bay deltas. The question as to whether it is desirable or practicable to recognize young, mature, and old stages of development of each of these particular forms is discussed, as is also the question as to which marine forces are principally concerned in their construction. The various forms discussed are illustrated by ideal diagrams and by maps of examples taken from nature.

Initial Stage. — As was early hinted by de la Beche[1], and later more clearly stated by Dana[2], when a land mass is submerged the sea enters the main river valleys and their lower tributaries for a distance which depends upon the depth of submergence, comes to rest against the more or less steep slopes of adjacent hills or mountains, and overflows the lowest cols or passes which separate outlying hills from the higher main ridges or divides. The initial stage (Fig. 45) of the typical shoreline of submergence is therefore characterized by an exceedingly *irregular shoreline,* many times longer than a straight line connecting two points on the shore; by numerous branching bays or *drowned valleys* in which comparatively deep water is found a short distance offshore; by many *peninsulas* projecting out to sea; by the presence of numerous *islands;* and by an *irregular sea-bottom* whose inequalities represent the former hills and valleys of the land. Portions of the coast of Maine and of the

Chesapeake Bay region (Fig. 46) are bordered by shorelines of submergence, but little changed from the initial form, outside of which submarine hills and valleys are clearly shown by soundings.

The great variety of form which initial shorelines of submergence may possess has already been suggested by the clas-

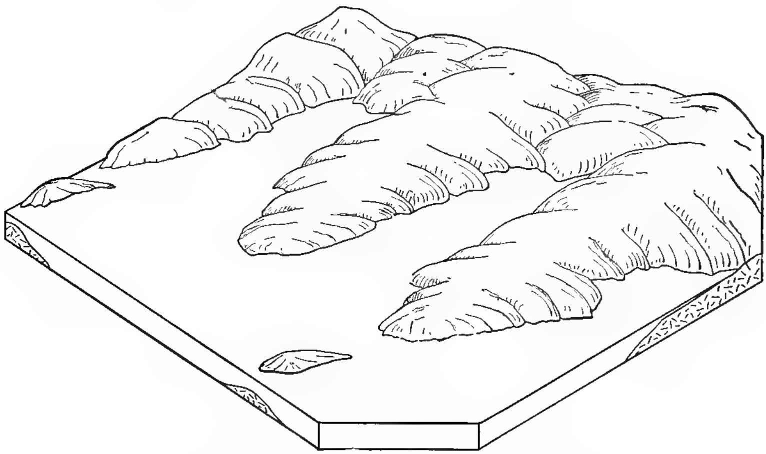

Fig. 45. — Shoreline of submergence, initial stage.

sification of shorelines outlined in Chapter IV. Such variety is inevitable, as will readily appear when we consider that everything which affects the shape of the land must also affect the form of the shoreline produced when the sea surface comes to rest against the land. A land mass may have a great variety of structures, which will be reflected in the shore forms. Those structures may be subjected to several different erosive processes, each of which produces surface forms peculiar to itself, and hence leaves its impress upon the shoreline of submergence. Each erosive process may be in any stage of its cycle when submergence occurs, and the resulting shore features will vary widely with the different stages of land form development. The stage of shoreline development reached at any given moment since submergence will, of course, profoundly affect the characteristics of the shore.

It is essential, therefore, to a clear conception of the characteristic features of any shoreline that the description take account of the structure of the land mass, the process or processes

FIG. 46 — Shoreline of submergence in Chesapeake Bay region. Except on the more exposed outer coast the shoreline is not greatly changed from its initial form.

by which the land mass has been eroded, the stage of land mass dissection reached when submergence occurred, and the stage of shoreline development reached since submergence. To say that "the coast of Dalmatia represents a region of folded mountains, maturely dissected into longitudinal ridges and valleys by normal stream erosion, and then slightly depressed to form a shoreline of submergence which is now in the youthful stage of its development," will bring to the hearer who is familiar with the elementary principles of shoreline development a clearer mental picture of the essential characteristics of that shoreline than could a much longer and more detailed account of individual bays, peninsulas, islands, and other local features. Greater definiteness may be given to the mental picture if the explanatory description quoted above is made to include a statement

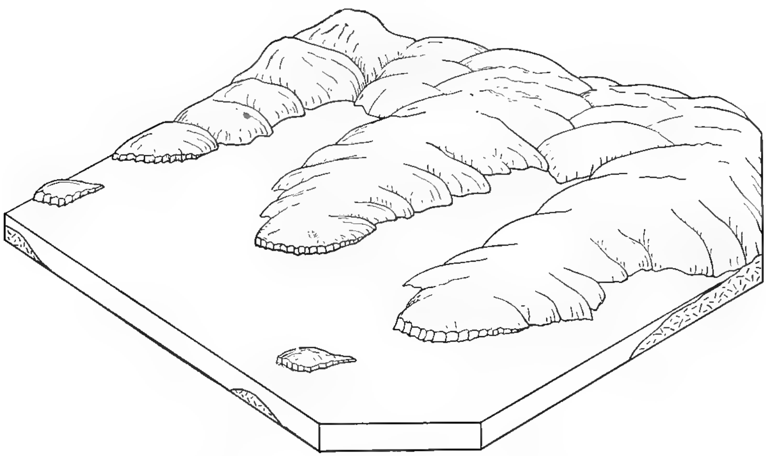

FIG. 47. — Early youth of a shoreline of submergence, showing crenulate shoreline.

as to the relief and texture of the topography produced by stream erosion; for the coast will be bold or subdued according as the relief is high or low, and the bays will branch moderately or intricately according as the texture is coarse or fine.

Young Stage. — As rapidly as submergence brings the hill and valley slopes within reach of the sea, waves attack those slopes. We have already seen that in the early stages of wave attack the cliff profile is more irregular than in the initial stage,

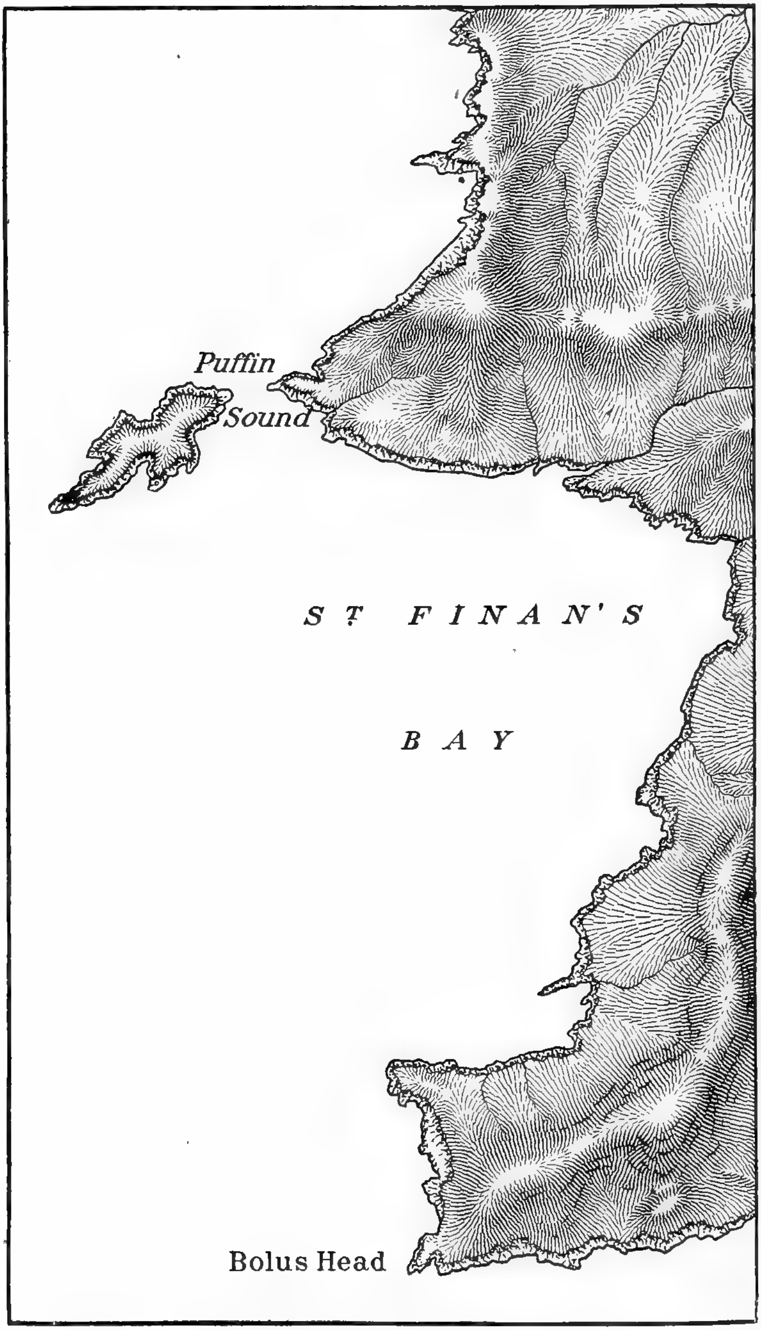

FIG. 48. — Crenulate shoreline of the southwest coast of Ireland.

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PLATE XXXIV.

Stack or chimney in front of a young cliff on the coast of France.

because resistant and non-resistant rocks are unequally affected by wave erosion. In a similar manner the initial shoreline is rapidly made extremely irregular, on a small scale, wherever the land presents to the sea rocks of unequal resistance. The hills and valleys of the land may have been well graded and characterized by smooth, flowing contours, in which case the initial shoreline must be composed of well-rounded curves. But early in the youth of the shoreline the curves will be changed to sharply and irregularly crenulate lines by differential wave erosion[3] (Fig. 47). In other words, although the ultimate goal of wave erosion is to make a shoreline of submergence less irregular, as will presently appear, the first effect is to make it minutely more irregular. We may call a shoreline of this character a *crenulate shoreline*. The shoreline of southwestern Ireland (Fig. 48), bordering the beautifully graded hills of a maturely dissected and partially submerged land area, is in this crenulate stage of development, as may readily be observed from the deck of a transatlantic steamer passing near the coast on its way to Liverpool. Portions of the coast of Japan (Fig. 49) likewise afford excellent examples of crenulate shorelines.

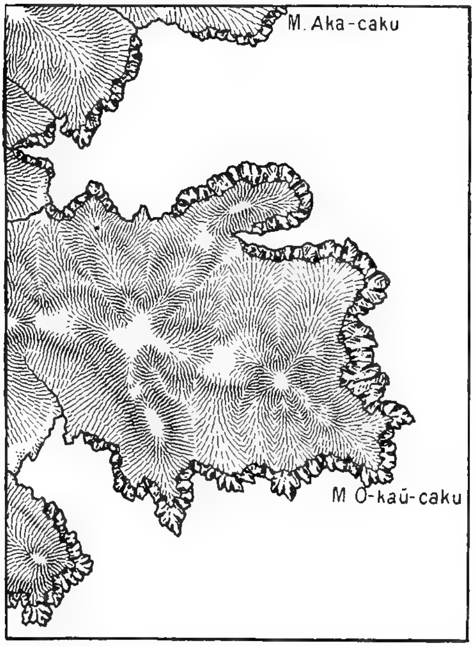
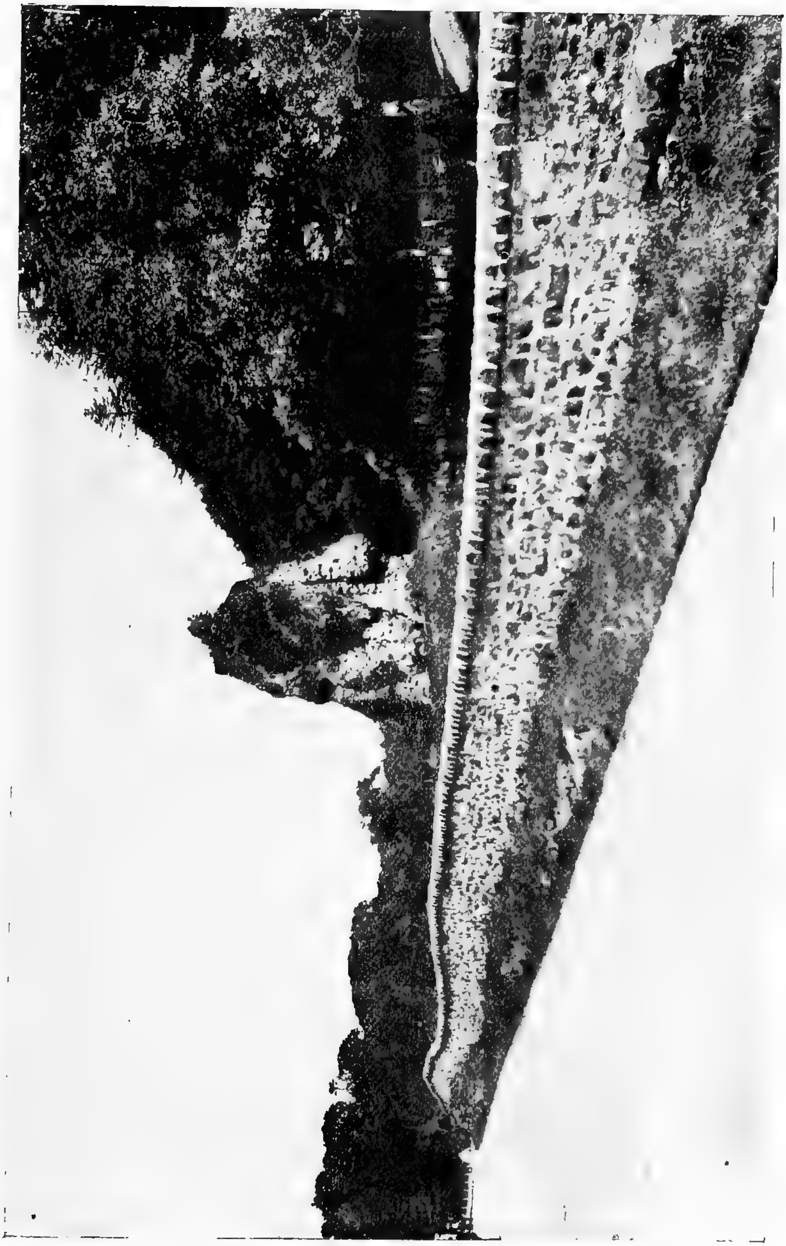


FIG. 49. — Young shoreline of submergence near Idzuhara, Japan, showing crenulate stage. (From Russian map based on Japanese data.)

During early youth some of the most picturesque features of cliff detail begin to appear. On rocky shores isolated pinnacles of resistant material are left standing for a time in front of

PLATE XXXV.



The Dogstone, near Oban, Scotland, a stack in front of a marine cliff, now elevated 25 feet above sealevel.

PLATE XXXVI.

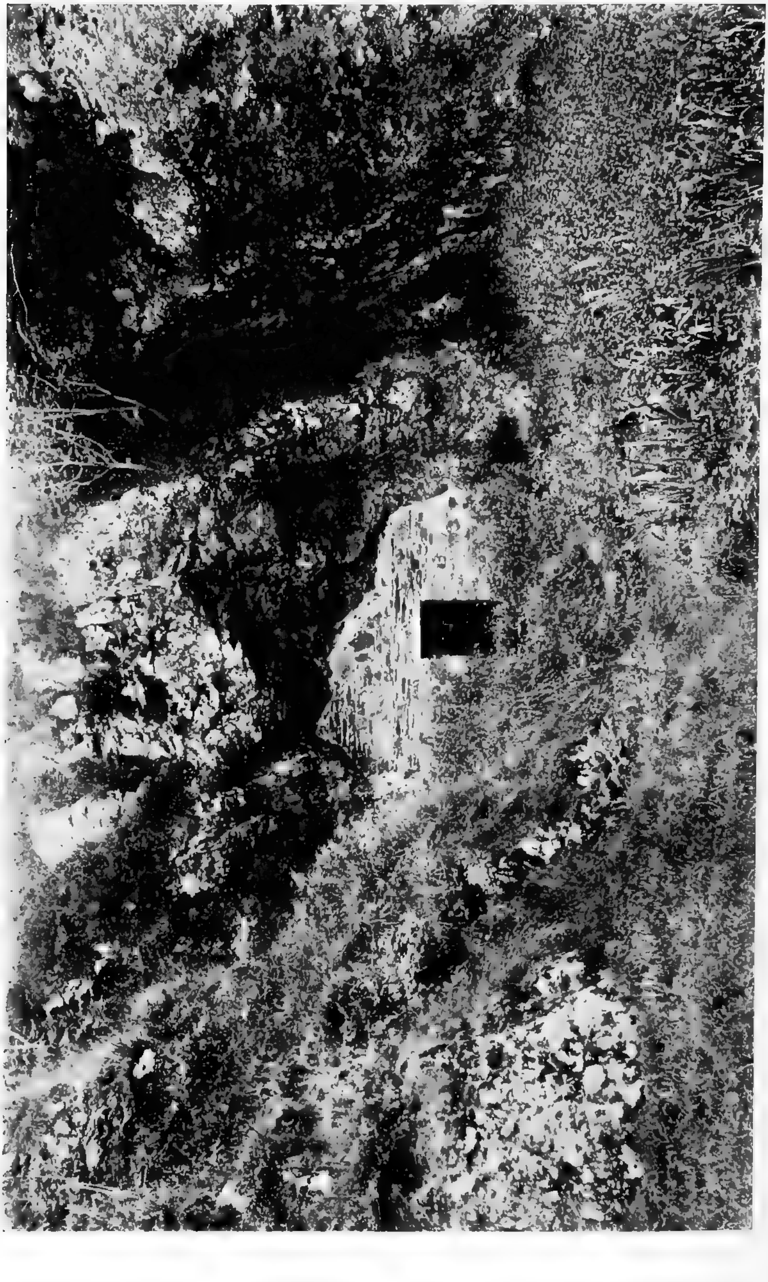
Fingal's Cave on the island of Staffa, Scotland. A sea cave formed by wave erosion in columnar basalt.

the main cliff. These *chimneys* or *stacks* (Plates XXXIV and XXXV) may be sculptured by the waves into very striking forms. Weaker zones are excavated by the waves into *sea caves* (Plates XXXVI and XXXVII), of which Fingal's Cave on the island of Staffa is a well-known example. Where a projecting belt of rock is completely pierced by the wave attack, an *arch* (Plate XXXVIII) is formed. In front of the cliff low tide may expose a bare rock platform representing the landward edge of the marine bench upon which the occasional stacks are situated (Plates XXII and XXVIII). From the face of the cliff numerous *landslides* (Plate XXXIX), usually small but sometimes of grand dimensions, are precipitated into the water or upon the rock platform as a consequence of the rapid encroachment of the waves along the cliff base. It should be noted that while the above-named features begin to appear in the early youth of the shoreline of submergence, and reach their most abundant development before maturity is attained, they may also be present on fully mature shores.

Because of wave refraction, the seaward ends or "headlands" (Plate XXXIII) of peninsulas and islands are more vigorously attacked than other parts of the shore, while the inner ends of the bays, or "bay heads," suffer least. In a comparatively short time, therefore, there are developed *cliffed headlands* of striking aspect (Fig. 50, *ch*). Part of the material eroded from the headlands is deposited in the depressions of the irregular seafloor, a second part is carried out to the deep sea, while a third part is temporarily built into various types of beaches and embankments.

The great variety of forms assumed by these beaches and embankments is dependent upon the unorganized condition[4] of the longshore currents near a young shoreline of submergence, and distinguishes the latter from all other classes and stages of shorelines, which are much more simple. As shown by Figure 51, tidal currents are broken up and deflected in various directions by the sinuosities of peninsulas, islands, and drowned valleys, whenever they impinge upon an irregular coast. Beach drifting under the influence of the swell and of direct wind waves will be equally irregular, and will often be opposed to the direction of tidal currents. The complexity will be increased wherever other types of currents are disintegrated against the irregular

Plate XXXVII.



Ancient sea cave on an elevated shoreline of western Scotland, transformed into a stable.

shore. Eddy currents are unusually numerous along such a coast. Wave-eroded débris which is moved by any of these currents must accordingly be built into an almost endless variety of isolated forms not intimately related to each other.

Beaches. — In early youth no very extensive beach is apt to form at the base of the headland cliffs, although narrow *headland beaches* (Fig. 50, *hb*) may be found in favored localities, especially if the cliff is composed of non-resistant sand or other material

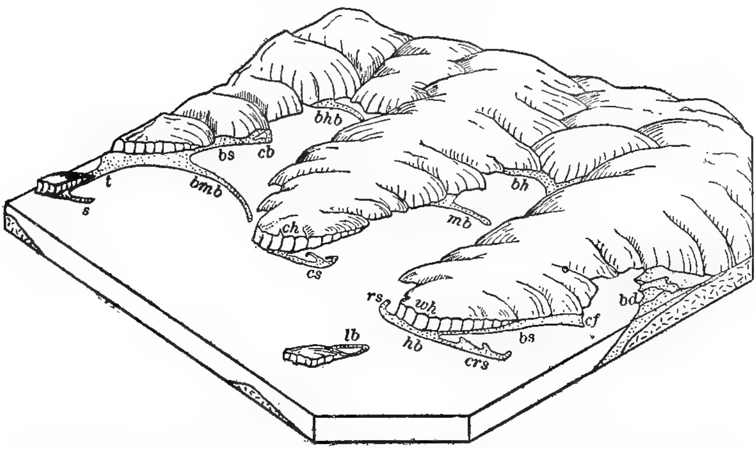


Fig. 50. — Young shoreline of submergence, showing types of beaches, bars, spits, and forelands.

which readily disintegrates. Most of the débris, however, is swept from the marine bench at the base of the exposed cliff as rapidly as erosion and weathering remove it from the cliff face. Beach drifting, possibly aided by other types of longshore current action, propel a considerable portion of the débris along the shores of the bays toward the bay heads. These latter areas are loci of deposition because wave refraction has here reduced wave erosion to a minimum, because beach drifting due to on-shore wind waves and to the swell is far more potent than any beach drifting which can result from offshore winds, because direct wind currents moving from the ocean surface into the bays are more effective than wind currents originating at the heads of the bays and moving seaward, and because flood-tide currents following the shores of a narrowing bay are apt to be more power-

PLATE XXXVIII.

Wave cut arch on the northwest coast of France. In the distance can be seen a stack and a second arch.

ful than the opposing ebb currents. It happens, therefore, that much of the débris eroded from the headlands is built into *bay-head beaches* (*bh*) at the inner ends of adjacent bays (Plate XL). The material in transit along the sides of the bay may form *bay-side beaches* (*bs*) which when fully developed connect the usually unimportant headland beaches with the more often well-developed bay-head beaches.

Embankments. — As may be observed from Figure 51, the shore currents of a young shoreline of submergence sometimes pass

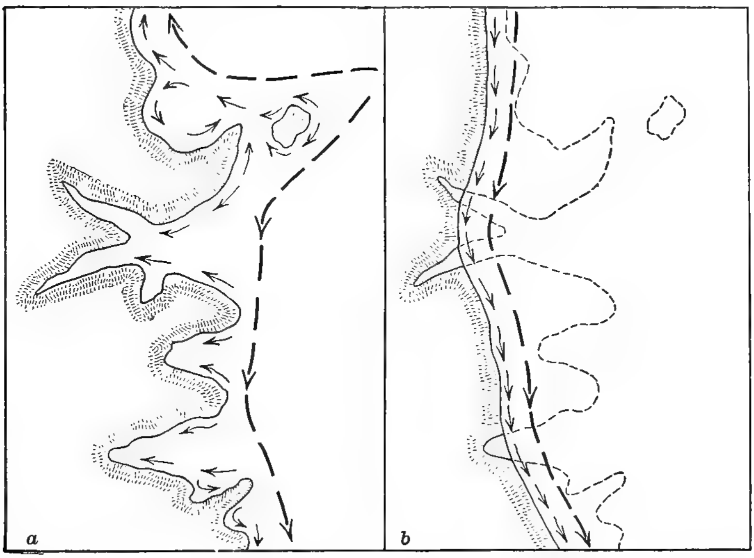


Fig. 51. — Initial unorganized condition of currents along a young shoreline of submergence (left hand figure) compared with organized condition which obtains when the stages of submaturity or maturity are reached (right hand figure). Light arrows = longshore currents, heavy arrows = offshore currents.

directly across the mouths of subsidiary bays instead of closely following the trend of the shore; or an offshore current, such as a planetary or large eddy current, may keep its course past the outer headlands but little influenced by the bays. Under these conditions the shore débris may be built out into the water in the form of a narrow embankment which grows by an excess of deposition at its seaward terminus, just as a railroad embankment is extended

PLATE XXXIX.

Third Cliff near Scituate, Massachusetts, showing small landslide due to wave erosion of cliff base.

by the dumping of car-loads of débris at its free end[5]. In the case of the current-built embankment, deposition takes place partly because the current, which is comparatively swift where it impinges against the headland or the already completed portion of the embankment and therefore able to transport a large amount of débris, loses part of its velocity when it passes into the deeper, open water off the bay mouth; and partly because the débris, as soon as it reaches deeper water, is no longer effectively agitated by normal wave action which in shallow water served to raise it intermittently into the moving water of the current. The seaward side of the narrow embankment is acted upon by the ocean waves, which build its crest above normal sea-level and establish a profile of equilibrium, similar to that of an ordinary beach (Fig. 36). The quiet-water side may have a more uniform slope, determined by the subaqueous angle of repose of the deposited material. If the débris is coarse, the distal end of the embankment will have an abrupt slope to deep water, which also represents the subaqueous angle of repose of the material composing the embankment; but if the débris is fine, deposition will be less sudden, and the distal end will slope more gradually into deep water.

Fig. 52. — Sand spits on the shore of Port Orchard, Washington.

Spits. — So long as an embankment has its distal end terminating in open water, it is called *spit* (Fig. 50, *s*. See also Figs. 52 and 53). When the spit first begins to develop the longshore current responsible for it is normally so effective in comparison with other currents that the latter have little or no effect upon its form. Tidal currents may pass in and out of a bay at right

PLATE XL.

Bay-head or pocket beach near Rye, New Hampshire.

angles to the spit's direction of advance, or beach drifting may, under the influence of onshore winds, tend to drive the débris at the terminus into the bay; but deposition by the dominant longshore current lengthens the spit so rapidly in the direction of that current's intention that the weak or intermittent efforts of contrary currents produce no sensible effect. With the continued growth of the spit, however, and the consequent narrowing of the entrance to the bay, tidal currents pass the end of the spit with an ever-increasing velocity. More and more of the débris brought by the longshore current is carried in toward the bay by the flood tide, thus giving a landward deflection to the embankment. Outflowing currents sweep some of the débris seaward, but the combined effects of the longshore current and active wave erosion normally prevent any marked seaward deflection. The longshore current may itself be deflected toward the bay by the flood tide, thus assisting in the landward deflection of the spit it is building. Furthermore, when the spit first begins to grow, its elongation proceeds with comparative rapidity, because the water is shallow and no great amount of débris is necessary to build the embankment up to the surface. But as it advances into deeper water, more and more of the débris must be laid down in the depths, and less and less is available for the linear extension of the spit. Under the new conditions of slow advance the influence of flood tide and of beach drifting due to onshore winds becomes increasingly apparent, the débris at the terminus is

FIG. 53. — Simple spit (below) and compound recurved spit (above) at entrance to Port Moller, Alaska.

carried farther landward before new supplies are laid down in front of it, and a landward deflection of the spit results. Under these and other similar conditions it often happens that the end of a spit is more or less strongly curved inward. When the growing embankment acquires this form it is called a hooked spit, or better, a *recurved spit* (Fig. 50, *rs*).

The forces supplying débris to the longshore current, the longshore current itself, and the contrary currents which tend to recurve the spit, do not always act with even approximate uniformity. One or more of these activities may have a very pronounced intermittent character. In such a case, the forces tending to elongate the spit in a straight or slightly curved line may prevail for a period, after which the forces operating to recurve the spit may temporarily gain the ascendancy. The effect of this intermittent action will be to produce a spit whose inner side is diversified by a series of landward deflected points representing successive recurved termini. To this interesting form the name *compound recurved spit* (Fig. 50, *crs*) may be applied.

It sometimes happens that after a recurved spit is formed, new currents arise which remove material eroded from the more protected parts (usually the inner side) of the spit, and build it into a new embankment which is really essentially independent of the form from which it projects. The original and secondary spits do not curve or merge into each other; on the contrary their lines of growth intersect at distinct angles, indicating their independent relationship. The secondary spit is no more an integral part of the original spit than the latter is an integral part of the cliffed headland from which it springs. Since it is desirable to give the combined spits a single name because of their association in nature, we may speak of the grouped features as a *complex spit* (Fig. 50, *cs*). Sandy Hook is an excellent example of a compound and complex recurved spit. The landward curvature of successive termini is clearly indicated by the contours on the Sandy Hook topographic quadrangle. But the southwardly deflected embankments which have generally been regarded as representing merely an extreme amount of recurving of the original spit are seen on closer examination to be independent secondary spits built by the waves and currents of Sandy Hook Bay with material eroded

from the northwesterly trending embankments of the original form (Fig. 57).

Occasionally it happens that variable or periodically shifting currents extend a spit first in one direction, then in another, giving to it a more or less serpentine pattern. To this comparatively rare form the name *serpentine spit* may be applied. Gulliver[6] mentions two examples of this type in his essay on "Shoreline Topography."

As the cliff from which a spit springs is cut back by the waves,

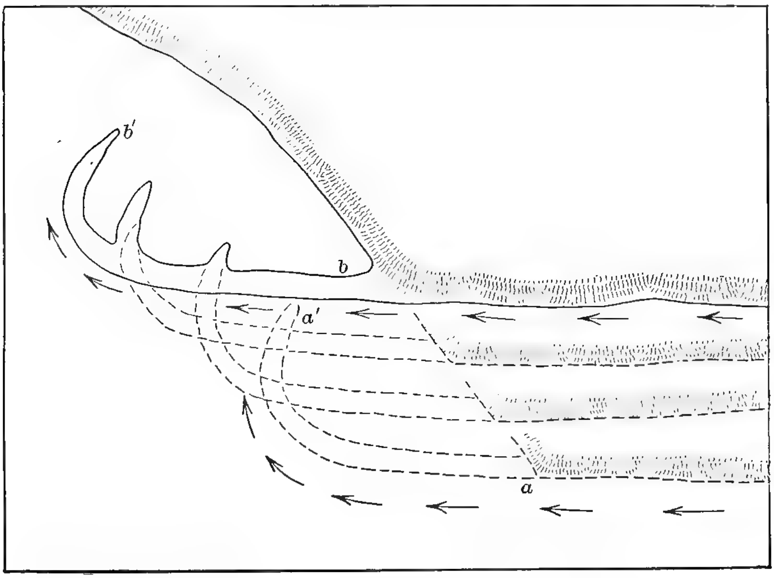


FIG. 54. — Successive stages in the development of one type of compound recurved spit.

the spit itself is driven landward at the same rate. This retreat of the shore leads to several interesting results, as will readily appear from Figure 54. If *aa'* is the original position of the spit, and the cutting back of the cliff causes the point of its tangency with the mainland to migrate toward the left, then the spit will assume the several positions indicated until it arrives at *bb'*. As will be observed, the spit may thus acquire a compound form with a succession of recurved points on its inner side, without necessarily experiencing any extension of its absolute

length. It should be noted, furthermore, that the progressive increase in the length of the ancient points does not in this case indicate a progressive increase in the relative strength of the landward moving currents, as has been generally assumed in such cases.

The marked angle at which the present shore intersects the axes of the ancient recurved points is evidence of their genetic relation to a former seaward position of the spit. This angle is normally greatest in the case of the oldest points, and decreases progressively in those which are of more recent date; but increasing effectiveness of landward moving currents may sometimes cause the latest points to bend landward at increasingly greater angles. Should the spit increase in length at the same time that it is pushed back, we may have a case in which the compound feature is observable only in the distal portion, the landward end being a simple, straight embankment (Fig. 55). This does not mean that the two unlike parts of the spit have had different histories, but merely that the landward end has been pushed completely back of the termini of the recurved points which formerly existed seaward of it.

The characteristics of a retreating compound spit are well seen in the embankment which encloses the harbor of Toronto on Lake Ontario. Figure 55, based on charts by Hind,[7] exhibits a compound distal portion, where an admirable series of recurved points are separated by subparallel ponds or lagoons, and a simple landward portion which has been pushed back beyond the position of the corresponding points in that region. The greater length of the remaining portions of the more recently formed points, and the greater angle which the oldest points make with the present shore, are well shown. Judging from the charts the ends of the recurved points are truncated by subordinate spits, which give the whole a more complex form than it would otherwise have. In 1854 Sanford Fleming[8] published an excellent essay on the history of this interesting shore form, clearly setting forth the essential stages of its development previous to the time of his study, and predicting probable future changes.

In the case just mentioned both the mainland cliff and the spit have retreated landward. There are cases in which the cliff beyond the base of the spit is cut back much more rapidly

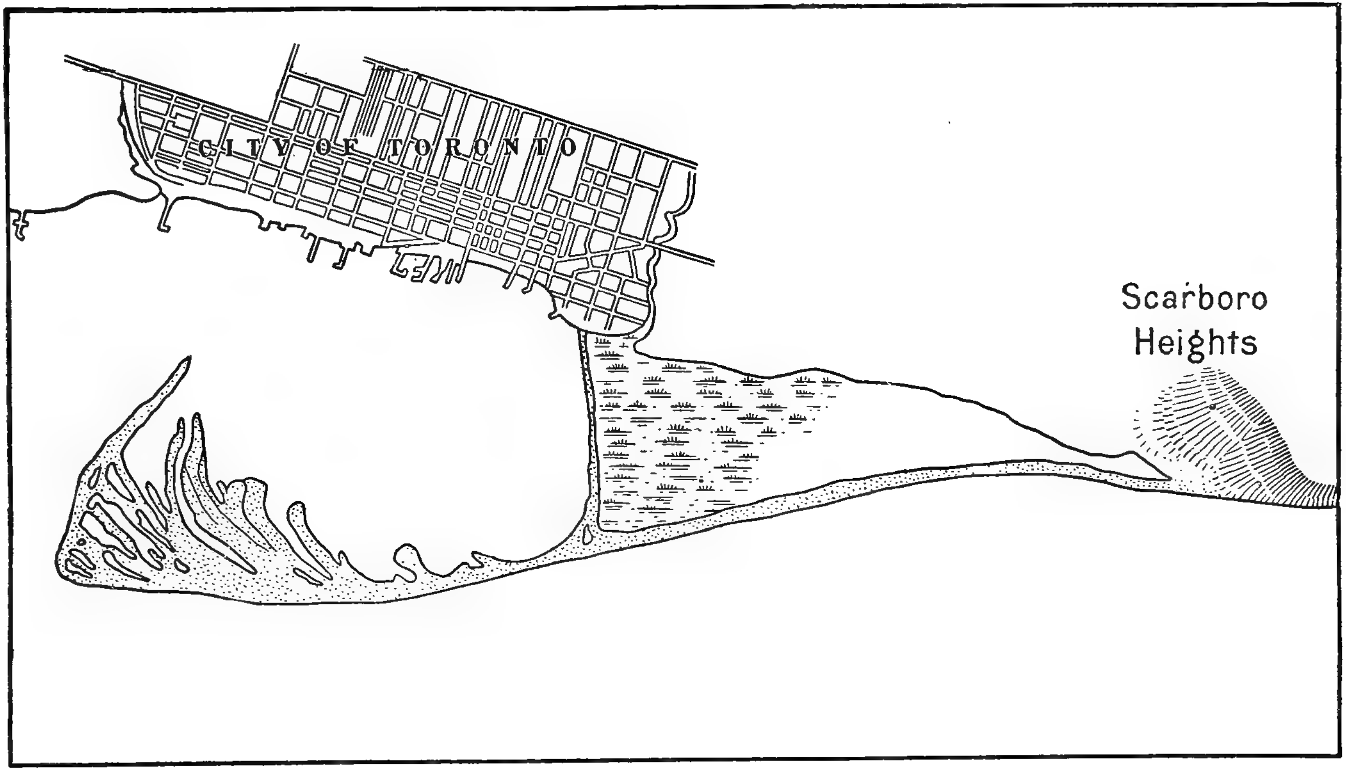


Fig. 55. — Compound recurved spit enclosing Toronto Harbor.

than at the point of attachment, while the spit as a whole advances seaward instead of retreating. Thus, in Figure 56, where a former spit (E) springs from the cliff base at the point B, the cliff southeast of this point is cut back very rapidly, while to the northwest there is little or no cliff erosion and the mainland is being protected by the growing spit. The cutting back of the

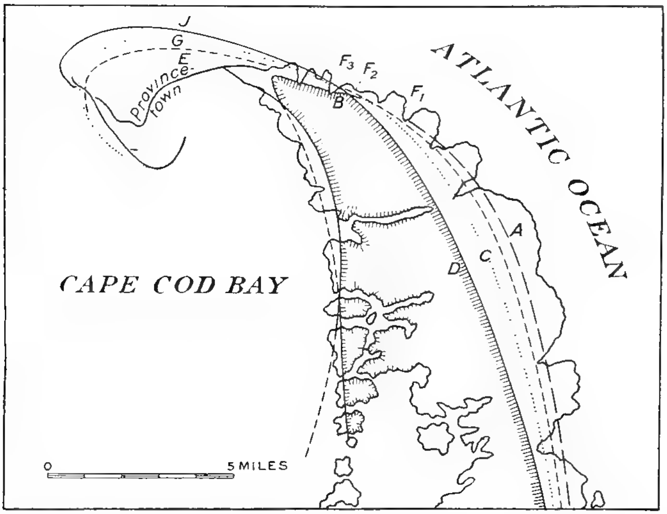


Fig. 56. — Development of the Cape Cod shoreline (after Davis). As the shore facing the Atlantic Ocean is cut back toward the west (A, C, D) the Provincetown sand spit grows progressively seaward (E, G, J), and the fulcrum point near B shifts from F^1 to F^3.

cliff at the southeast produces two important results: First, the direction of the shoreline is changed, so that the longshore current responsible for the spit comes from a more southerly direction and hence tends to maintain its course more toward the north after it passes B; in the second place, the cutting back of the shore removes the many irregularities which formerly disintegrated the currents impinging upon them. a single vigorous current along the simple shoreline replaces the many weak ones which flowed here and there along the complex shoreline, and this more vigorous current will maintain its more northerly course into open water after passing the point B because there is no

force competent to deflect it into the bay as readily as the original weaker currents were deflected. As a result of these conditions, the axis of spit-building moves progressively seaward, as the cliff to the south is pushed progressively landward. Near the point *B* there is a "fulcrum" (*F*), north of which the shoreline is everywhere prograded, while south of it there is only retrograding. As Davis has shown in his classic essay on "The Outline of Cape Cod,"[9] which contains the first adequate presentation of the fulcrum idea, the fact that some erosion is experienced at the point *B* and on the adjoining base of the spit causes the fulcrum point to shift slightly in the direction of the spit from F^2 to F^3 in Fig. 56. (See also Fig. 57). The Provincelands of Cape Cod and Sandy Hook are both good examples of spits formed in the manner above described. In the case of Sandy Hook the position of the earliest part of the spit, corresponding to *E* of Figure 56, may be indicated by the low sandy beach plain on the northeast side of Navesink Highlands, while Island Beach represents the remnant of a later addition to the spit. Sandy Hook itself advanced to the north and east by the successive additions of recurved points, as the shore near Long Branch was driven back toward the west. The fact that the base of Sandy Hook spit connects with a bay bar at the present time, instead of with the cliff on the east end of the Highlands, introduces a slight complication.

Johnson and Reed have shown that in so complex a series of spits and bars as that composing Nantasket Beach on the Massachusetts coast, the phenomenon of a shifting fulcrum between a retrograding cliff and a prograding beach plain may occupy an important place in the history of the shoreline[10].

It has already been shown that the distal portion of a spit, and consequently of each recurved point representing a former distal portion, is submerged, the end of the embankment sloping down into deep water either abruptly or gradually according to the nature of the débris of which it is constructed. The super-aqueous portion owes its height primarily to the waves, but in the case of sand spits wind action may locally raise the level a number of feet by forming dunes. Disregarding the disturbing effect of the wind, the height of a spit will depend upon the exposure to wave action; big waves will cast the débris

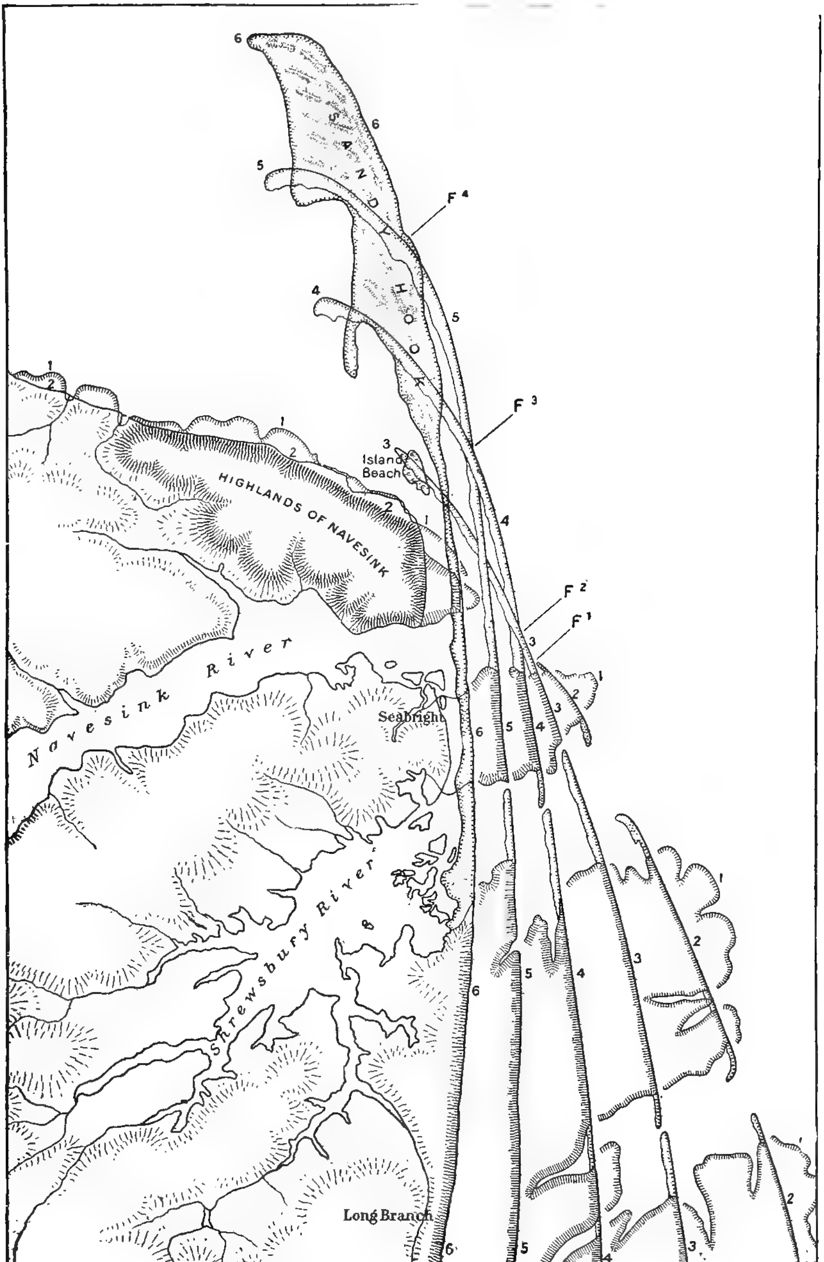


FIG. 57. — Development of Sandy Hook spit. As the original shore between Seabright and Long Branch was cut back by wave attack, the zone of spit formation north of Navesink Highlands advanced toward the northeast. The fulcrum point, dividing the zone of retrograding shoreline from that of prograding shoreline, shifted progressively from F^1 to F^4. West of the letters "oo" in "Hook" is a small southward-pointing spit built by waves from the northwest out of material eroded from the recurved points of the main spit.

many feet above mean water level, while small waves will raise the surface but slightly above the lake or sea. Since the exposure of a recurved spit to wave action is unequal, it follows that some portions must be higher than others; and it is normally the case that the distal curved portion, which is acted upon by the smaller waves of the bay into which it is being deflected, has a distinctly lower crest line than the rest of the spit. The importance of a proper appreciation of this simple relation will appear when one remembers that the low altitude of the crests of the recurved points in a compound spit have erroneously been regarded by some observers as a proof of coastal subsidence.

The successive embankments added to a growing compound spit may be closely spaced, with shallow depressions between them whose bottoms do not extend as low as sealevel; or the embankments may be widely spaced and separated by lagoons of fairly deep water. If the supply of débris, longshore current action, and the activity of other currents are fairly uniform and constant, the successive embankments will be closely spaced and tend to form a continuous plain, which we may call a *beach plain* in view of the fact that it is composed of beach deposits cast up by the waves. It very seldom happens that all forces operating at the shore are so uniform and continuous as to give a perfectly smooth plain surface; on the contrary, the surface of the beach plain ordinarily shows a series of low ridges representing the crests of beaches built by the waves along successive positions of the shoreline. These *beach ridges*, or "fulls" as the English geologists call them, constitute lines of growth of the beach plain, and when well preserved enable one to trace the history of development with great accuracy. They vary in altitude according to exposure to wave attack, but from three to twenty feet above ordinary high water level may be taken as the more common elevations. Beach ridges are conspicuous features of certain other coastal forms besides spits, and will be further considered when those forms are described.

If any one or more of the forces involved in spit building operate very irregularly or intermittently, it may happen that successive embankments will be built at wide intervals. Let us imagine that the longshore current runs much more swiftly at rare intervals than at other times. For a long period it may

flow too slowly to remove all of the débris eroded from the cliff, and a large beach deposit accumulates at the cliff base and along adjacent parts of the spit. During this time such material as the current does transport is easily carried around the recurved point and in toward the bay, because the landward directed currents are competent either to move the load of a comparatively weak longshore current back to the previously established shoreline, or to deflect the longshore current itself so that it deposits its load directly along that shoreline. Now let us suppose that the longshore current is accelerated. Its increased velocity will enable it to pick up and transport the large load of débris which accumulated during its period of sluggishness; and will also impel the current to maintain its course straight ahead into deep water, instead of suffering deflection into the bay when it reaches the recurved point of the spit. Furthermore, the shore below the fulcrum has been cut back to some extent during the period since the last recurved point was formed; and while the effect of this was not readily apparent so long as the longshore current was comparatively sluggish, a vigorous current finds at once that the prolongation of its normal course lies to seaward of the distal part of the spit. Large quantities of débris, borne by a current which departs from the former shoreline and advances into open water, must be built into an embankment which elongates rapidly in the direction of current advance. Waves raise the surface of the new embankment into a beach ridge; and by repetitions of this process there are formed successive beach ridges separated by lagoons of considerable breadth. Irregular or intermittent activity of other shore processes may produce the same result.

Ordinarily the intermittent character of shore activities is not sufficiently pronounced to cause the building of new embankments so far removed from the older ones as to have really deep water between them. Usually a shallow lagoon or merely a marshy swale separates the ridges. The compound recurved spit at Toronto has a series of elongated ponds of shallow depth (Fig. 55), as has also the compound recurved spit known as Presque Isle on the south shore of Lake Erie (Fig. 58). Sandy Hook exhibits close-set ridges, or ridges separated by shallow, marshy swales. It is possible that in the latter spit the channels northeast and southwest of Island Beach (Fig. 57) occupy the

positions of lagoons or bays between former ridges now largely destroyed.

The ultimate length of a spit is attained when the tendency of longshore transportation to increase that length is just balanced by the opposite tendency of contrary currents. Where

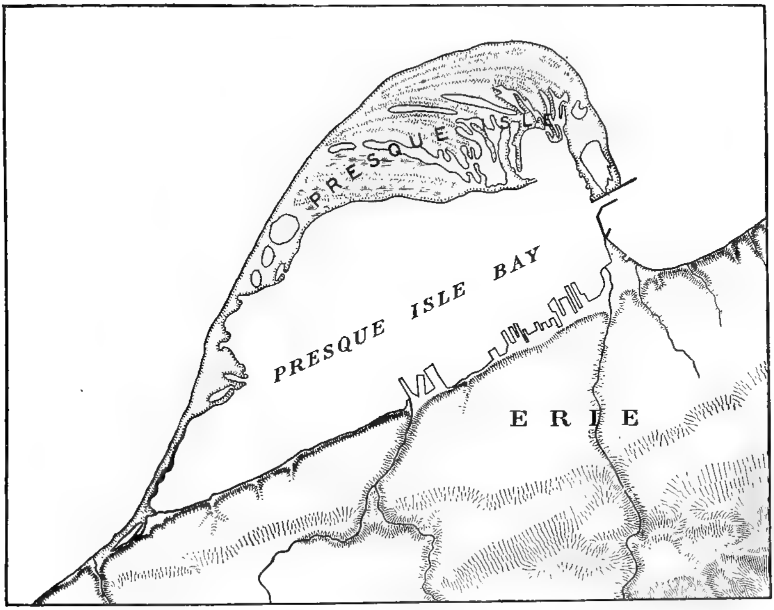


FIG. 58. — Lagoons and ridges of the Presque Isle compound recurved spit

the embankment is extending across a bay, progressive lengthening narrows the inlet through which tidal, hydraulic, and other currents must pass in and out, and thus increases the velocity of those currents. This process continues until the velocity of the latter currents is just high enough to counteract the constructive tendency of the longshore currents, when the embankment ceases to grow. The point of equilibrium is the sooner reached because wave action on the seaward side of the embankment continually reduces the size of the particles in transit, with the result that the farther the spit advances into open water the less powerful are the cross currents required to remove material from its distal end. Recurved points build into the bay until a similar condition of equilibrium is established be-

tween the currents tending to lengthen and those tending to remove the point. Because of the varying intensity of all shore processes, the equilibrium is never perfect, but only approximate; and the end of the embankment therefore advances and retreats intermittently over a narrow zone which might be called the "zone of equilibrium." It is believed by some that Sandy Hook has reached this zone of equilibrium, and that the currents into and out of New York Bay are now sufficiently strong to overcome the efforts of the northward flowing longshore current to increase the length of the spit. This view is expressed by Duane[11] in the following words: "It (Sandy Hook spit) appears to have reached a limiting length at which the currents into and out of New York Bay have sufficient strength to scour away sand deposited at its northern end, and in the last one hundred and forty-five years its length has varied only about 2700 feet, sometimes increasing and sometimes decreasing."

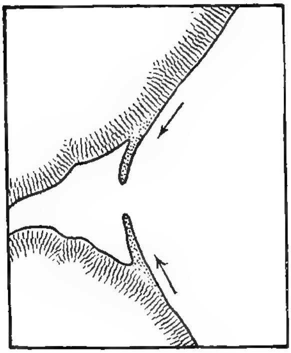


Fig. 59.

Bay Bars. — If the zone of equilibrium is not reached by the embankment until it has almost closed the inlet, or if the longshore currents prevail throughout and succeed in extending the embankment completely across the bay, the spit becomes a *bay bar*. A spit may thus change to a bay bar interrupted by a narrow inlet, and this in turn to an unbroken bay bar. Within a bay converging currents may build two spits toward each other until they form a bar (Figs. 59 and 60). As a rule the sea tends to build bars which are slightly concave toward the open water; or to drive back the central part of a bar more than the terminal portions until such concavity results. But where an embank-

Fig. 60. — Spits converging to form a bay bar on the Alaskan coast.

ment grows across the mouth of a narrow bay, and longshore current action is very powerful, the resulting bar may be quite straight.

There is, however, an entirely different process by which bars, indistinguishable in surface form from those developed from growing spits, may be produced. Waves entering shallowing water may break before reaching the coast, and cast up the bottom débris into a narrow ridge, in the manner discussed more fully in connection with "Offshore Bars." The irregular bottom of a typical young shoreline of submergence is usually highly unfavorable to this process; but whenever the initial form or later deposition does give a fairly uniform slope to the

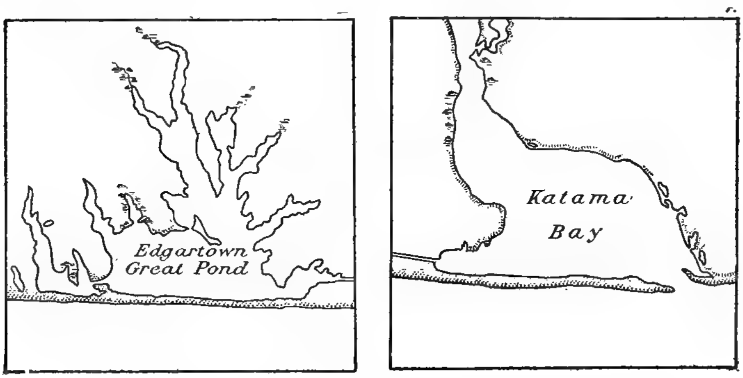


FIG. 61. — Bay-mouth bars on the Marthas Vineyard coast.

bottom near the shore, wave action may produce a bar independently of longshore transportation. Such a bar may form a short distance offshore and be driven in until the portion opposite a headland becomes a headland beach, and the portion opposite the bay remains a typical bay bar extending from headland to headland and nearly or quite closing the bay mouth; or the waves may construct the bar just at the mouth of the bay in the first place; or they may break on the gently sloping bottom well within the bay and produce a bar near the middle or even near the head of the bay. It is possible that some supposed sandspits are really the beginnings of, or last remnants of, bars formed in this manner.

A compound shoreline, like that of northern New Jersey, is

especially apt to have an offshore bar pushed landward against projecting headlands, after which it will appear as a series of shorter bay bars. The bar across the mouths of the Shrewsbury and Navesink Rivers (Fig. 57) may have had an earlier existence as an offshore bar farther out in the Atlantic; but its history is not altogether simple, for it has been temporarily breached, and later rebuilt in part at least by longshore currents. The same is true of the bay bars closing Shark River and Manasquan River, which probably originated as parts of one offshore bar; while Metedeconk River will in the future be closed by the northern part of the offshore bar on which the town of Mantoloking is situated.

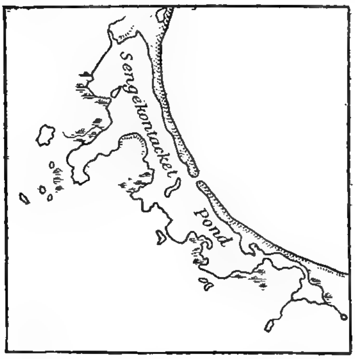


Fig. 62. — Bay-mouth bar on the Marthas Vineyard coast.

It does not seem desirable to give separate names to bay bars formed in the two ways above described, because of the fact that the method of origin is often obscure. There are cases, of course, in which the process of formation may be inferred with reasonable assurance from the form or position of the bar; as for example when successive recurved points on the inner side of a bar indicate its development from a compound recurved spit, or the abutting of the bar at right angles against the shores of the bay show the predominant action of waves breaking on a shelving bottom. It is probably true, however, that in a majority of the cases where offshore wave action originates a bay bar, longshore transportation plays an important part in its further development. To determine the relative importance of the two co-operating forces may well be impossible. We will therefore name bay bars according to their position in the bays across which they have been extended, admitting the existence of two processes which may independently or in co-operation produce them. On this basis we may recognize (1) *bay-mouth bars* (Fig. 50, *bmb*) or those extending from headland to headland across the mouths of bays, excellent examples of which are found along the shores of Marthas Vineyard Island

(Figs. 61 and 62); (2) *bay-head bars* (*bhb*) or those built a short distance out from the shore at the heads of bays, like the outer bar near Duluth at the head of the westernmost bay of Lake Superior (Fig. 63); and (3) *mid-bay bars* (*mb*) or those built across a bay at some point between its mouth and its head, a

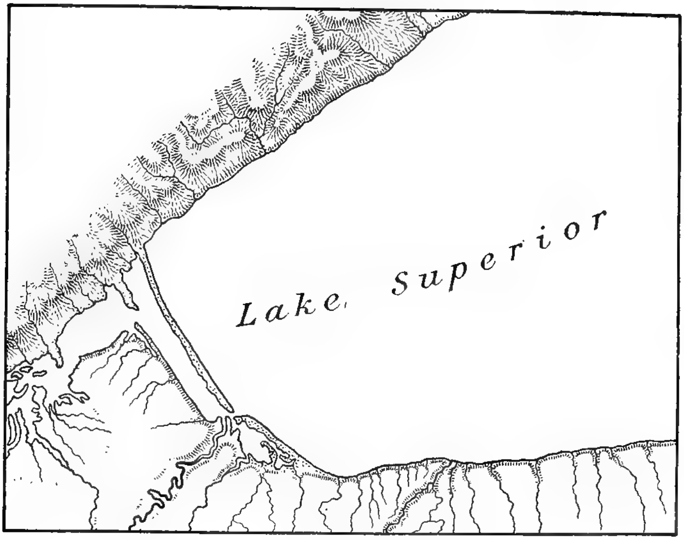


FIG. 63. — Bay-head bar near Duluth.

good example being the bar which extends nearly across the middle of Hempstead Harbor, Long Island (Fig. 64).

A headland which is bordered on either side by bay bars or spits is sometimes called a *winged headland* ("winged beheadland" of Gulliver[12]). Grassy Hollow headland near the eastern end of Long Island is a typical specimen of this interesting form (Fig. 65). At Long Branch on the New Jersey coast we have the very large winged headland which Gulliver selected as his type example.

After a bay bar has been constructed, the pond or lagoon enclosed behind it may gradually be transformed into a land area through the combined operation of several agencies. Streams from the land bring down sediment which may either be distributed over the floor of the lagoon by current action, or built into a bay delta which advances seaward until it meets

the bar. Tidal currents carry débris, from the zone of wave agitation outside the bar, through the inlet, and distribute it over the lagoon bottom or build it into a tidal delta (Fig. 117) which projects into the lagoon with its surface usually below

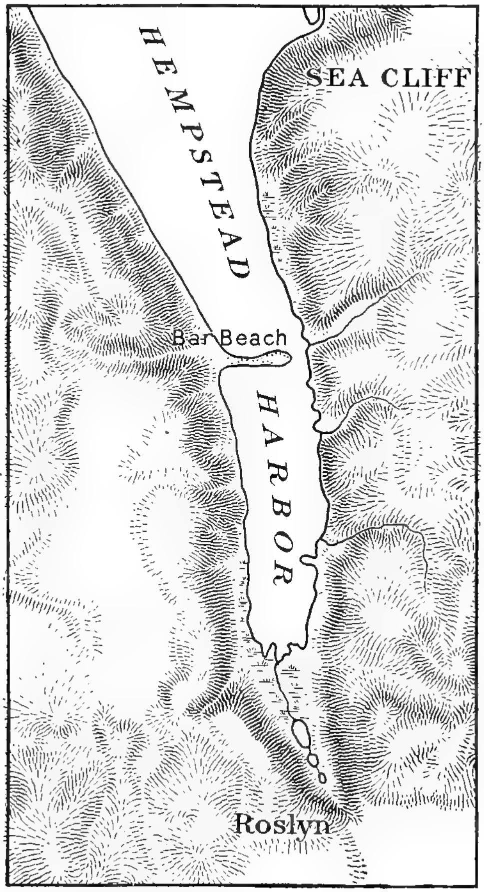


Fig. 64. — Mid-bay bar in Hempstead Harbor, Long Island.

sealevel. Winds from the sea blow sand from the surface of the bar into the lagoon behind it, and may even cause sand dunes to migrate some distance into the enclosed area of quiet water. Large storm waves dash over the crest of the bar,

PLATE XLI.

Deltas of cobblestone formed by overwash of storm waves near Marblehead, Massachusetts. The deltas are encroaching on the salt marsh back of the bar.

and their waters flowing down its landward side build *wave deltas* (Plate XLI) into the edge of the lagoon. Salt marsh vegetation may secure a foothold in the areas of shallower water, and both by building up to the surface and by advancing over

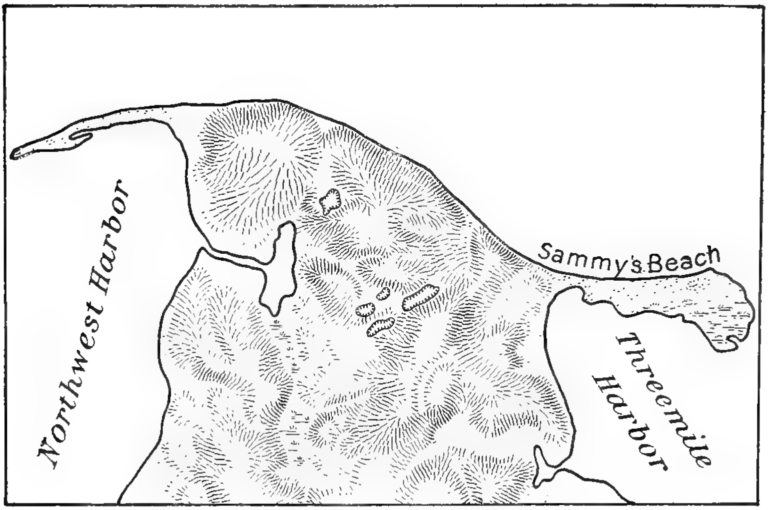


Fig. 65. — Winged headland near Sag Harbor, Long Island.

other portions of the lagoon may materially hasten the conversion of the entire area into land. The process of conversion goes on at very unequal rates in different places; it is essentially independent of the progress of shore development on the outer coast; and it may even depend mainly on forces which are not directly connected with marine agencies.

In the literature on shorelines one not infrequently encounters the curious idea that bay bars are the product of river deposition. The material of the bar is supposed to have been carried out to the mouth of the bay and dropped where the brackish bay water meets the salt water of the sea. Von Richthofen[13] seems thus to account for the bay bars of his " Liman type " of coast, the standard example of which is the northwest coast of the Black Sea near Odessa; and this theory is adopted by Hentzschel[14] and others in discussing the same and similar regions. How the coarse sand and even larger débris often composing the bar could be carried through the quiet waters of the

bay, and why such débris was not deposited in the form of a delta where the river enters the bay head are matters not satisfactorily explained.

Offsets, Overlaps, and Stream Deflection. — Where a bar has almost closed a bay mouth, a narrow *tidal inlet* maintains connection with the open ocean, and permits tidal currents to pass in and out of the lagoon. Rivers emptying into the lagoon may increase the outflowing currents; and where there is no tide the opening maintained principally by the outflow of river water alone might better be called an outlet, were it not that similarity of form and the desirability of uniformity in usage make it expedient to apply the single term "inlet" to all these features.

From the method of bar development it follows that an inlet is normally found at that end of a bar toward which the longshore current responsible for its growth is moving. It frequently happens, however, that storm waves break through a bar and establish an inlet at some other point, often at or near the point of attachment with the mainland cliff. Thereupon the original inlet may close, while the new one begins to migrate in the direction of the longshore current in consequence of the fact that deposition constantly occurs at the end of the bar on the up-current side of the inlet, necessitating an excess of erosion on the other side by the transverse currents which insist on keeping the inlet wide enough to permit their passage. In this manner the new inlet migrates until it reaches the position of the original inlet at the down-current end of the bar, when the process may be repeated. Sometimes the older inlet is closed before a new one is opened, and the bar exists for some time without any opening. Shaler[15] was of the opinion that new inlets were due to the bursting out of dammed-up land waters which had been held in restraint by an unbroken bar; but all the evidence available seems to show that even where tidal influence is unimportant, new openings are most frequently cut from the seaward side by the attack of storm waves. On the New Jersey coast inlets through the bars which obstruct the mouths of Manasquan, Shark, and other rivers or bays are constantly closing and opening, and migrate uniformly in the direction of the dominant longshore current.

The migrating of an inlet under the influence of a longshore

current is commonly accompanied by the development of features which may permit one to determine the direction of the current from accurate maps or charts. In many cases the part of the bar on the up-current side of the inlet is a little farther seaward than the part below the inlet, in which case the shore is said to be *offset* (Fig. 66, *a*). It is possible to have offsets where there is no inlet, as shown at *b* in the same figure. Very frequently a bar which offsets its remaining portion at an inlet also *overlaps* it as shown at *c*; and where a stream enters the sea without passing through a bay, the shifting of the inlet at the stream mouth may cause a pronounced *stream deflection* (Fig. 66, *d*). The longshore current or currents responsible for these features move from the outer toward the inner segments of the shore, or in the direction of stream deflection, as shown in the figure. Direct observation of shore currents is complicated by the fact that at the time of observation less important currents in an opposite direction may chance to prevail; but according to Gulliver[16], who first emphasized the importance of offset, overlap, and stream deflection as indicators of current movements, the direction of the *dominant* current is reliably indicated when one or more of the three features just mentioned is present.

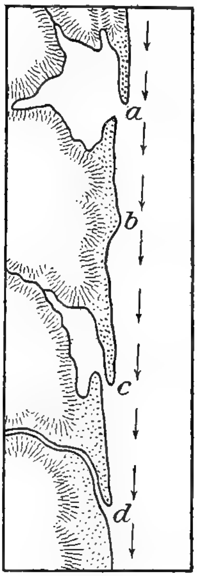


Fig. 66.

There is reason to believe, however, that direct wave attack may force one segment of a bar back of a neighboring segment, thus giving an offset which is quite independent of the direction of longshore currents. It might well happen that the resulting offset would be exactly opposed to that which would have been produced by current action. This seems to be the case on the southern part of the New Jersey shoreline, where the dominant current, as shown by the direction of inlet migration, is southward; yet successive offsets give a false indication of a northward moving current.

Stone Reefs. — Under special conditions ordinary bay bars or offshore bars may undergo a peculiar process of lithification which changes them into stone reefs. According to Branner[17],

who has described the remarkable series of stone reefs bordering the coast of Brazil for a distance of 1250 miles between Ceará and Porto Seguro, the only essential difference between these reefs and ordinary sand bars and spits lies in the induration of the upper ten or twelve feet of the sand through the cementing action of calcium carbonate. It appears that in and about the lagoons or ponds back of the bars abundant aquatic and semi-aquatic plants live and die. "The fresh water is thus rendered acid by the presence of large quantities of carbon dioxide produced by organic decomposition. The acid water on the land side percolating through the embankment of sand at low tide attacks the calcareous matter (fragments of shells, etc.) in the sand and passes seaward with it in solution, but as it comes in contact with the dense sea water on its way through the sand, the lime carbonate in solution is deposited in the interstices between the sand grains. In time the interstices are completely filled, and the sand bank is hardened and so solidified that the water can no longer soak through it." In Branner's opinion the essential conditions are the following: lagoons or ponds nearly or quite closed by bars or spits; abundant vegetation in or about these water bodies; fragments of shells, crinoids, coral, or other calcareous material in the bar or spit; and a high density of the sea-water. Stone reefs are rare because the combination of all these features is rare. Lithified beaches originally composed of sand and gravel and later cemented by calcium carbonate were early described by Beaufort[18] from the coasts of Asia Minor and Greece; while Cold[19] reports stone reefs from this same general region which separate lagoons from the open sea and which must be similar to those studied by Branner.

Looped Bars. — The islands of a young shoreline of submergence are attacked from all sides by the waves; but the most effective attack is delivered upon the seaward side, because both the swells and the largest storm waves come from the open sea, and because wave refraction concentrates the energy of both types of waves upon the seaward side of islands as well as upon projecting headlands. As in the case of headlands, part of the eroded débris is carried out to a permanent resting place in deep water, part is temporarily deposited in depressions of the irregular sea floor near the land, and part is built into various

types of beaches and embankments. Among the latter there are, in addition to the spits and bay bars already described, two forms peculiar to eroded islands: looped bars and tombolos.

Little beach material can accumulate at the base of the exposed cliff on the seaward side of the island. Sometimes spits extend out on either side of the main cliff, usually with their axes directed backward toward the land. More often, perhaps, the débris is driven backward along either side of the island until the quieter water to leeward is reached. Here embankments of several types may form. Spits may trail backward from either side, maintaining a separate existence; or their ends

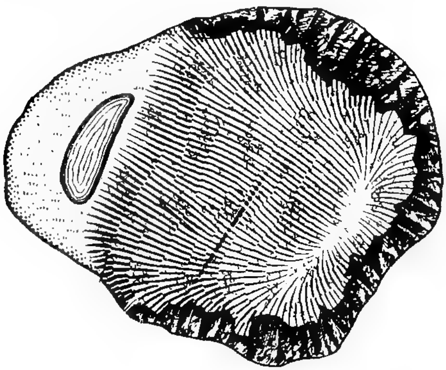


Fig. 67. — Looped bar on shore of Shapka Island, Alaska. (C. S. Chart, 8881.)

may unite to form a *looped bar* (Fig. 50, *lb*). Shapka Island, Alaska (Fig. 67), and Cup Butte[20], Utah, furnish good examples of looped bars, the latter existing as an elevated shore feature on a former shoreline of Lake Bonneville. In other cases one or more embankments will be extended until the island is directly connected with the mainland. The extension of the embankment may take place wholly in the direction from the island toward the mainland (Fig. 68); or wholly from the mainland toward the island, especially in those cases where longshore currents build a spit out laterally until it forms a bar which connects with an island lying to one side of the cliffed headland; or the embankment may be constructed from both directions at the same time until the ends meet to form a connecting bar; or, finally,

the bar may be built up simultaneously along its entire length by wave action on a shallowing bottom (Fig. 69). Furthermore, the mainland may be replaced in any of the above instances by another island, without altering the essential relations. In all of these cases the connecting bar is called a *tombolo* (Fig. 50, *t*).

Tombolos. — The name "tombolo" was applied to the connecting bar by Gulliver[21] in the following words: "Upon the coast of Italy where island-tying in its various stages is beautifully shown, such a bar is called a tombolo. For convenience in distinguishing island-tying bars from

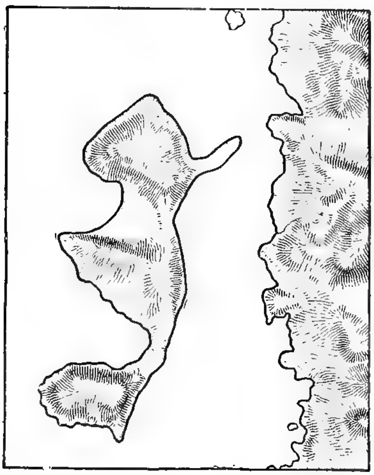


FIG. 68. — Renard Island near Seward, Alaska, showing embankment growing from island toward mainland.

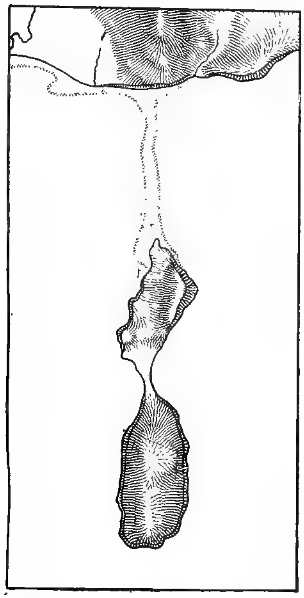


FIG. 69. — Inner Iliasik Island, Alaska, showing embankment which may be upbuilding toward the surface simultaneously along its entire length.

those of other kinds, the writer proposes to call every bar of this kind a *tombolo*, giving an English plural tombolos." Professors Olinto Marinelli of Florence and Giuseppe Ricchieri of Milan have both expressed to me orally their opinion that the term tombolo in the Italian language is restricted to the sand dunes found upon shore beaches and in other localities, and that it cannot properly be applied to a bar built by currents and waves. There seems to be no doubt that the plural "tomboli" does signify sand

dunes or similar small mounds. On the other hand, it would appear that failure of the popular mind to appreciate the independent origin of the bars and the dunes which surmount them, had resulted in the application of the term tomboli to the bars themselves, at least in some parts of Italy. This fact,

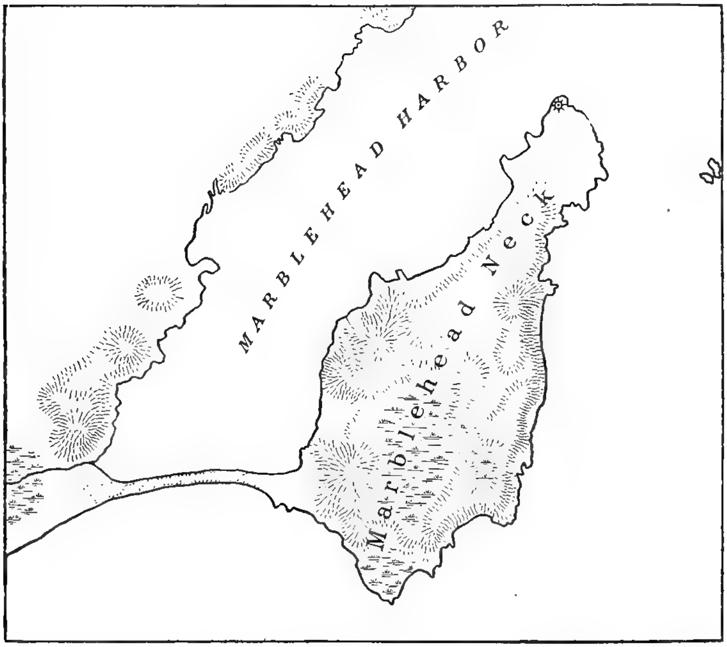


Fig. 70. — Single tombolo connecting former island of Marblehead with the mainland.

and the confusion of ideas responsible for it, are both shown in the following quotation from Pianigiani's Dizionario Etimologico della Lingua Italiana[22]: "'Tomboli' is a term commonly applied figuratively to the mounds of sands which the sea forms in the fashion of banks on the shore; otherwise called 'cotoni' = 'costoni' from 'costa:' for example, 'the sea, roughened by opposing currents or winds, scrapes the bottom and brings the sand back to the shore, forming *tumoli* or *tomboli*, and makes bars or shoals at the mouth of the Arno. These tomboli are the same thing as the famous dunes of the Dutch

and French' (Targioni, *Viaggi*)." * Prof. A. A. Livingston of Columbia University, to whom I am indebted for calling my attention to the foregoing citation, informs me that small mounds

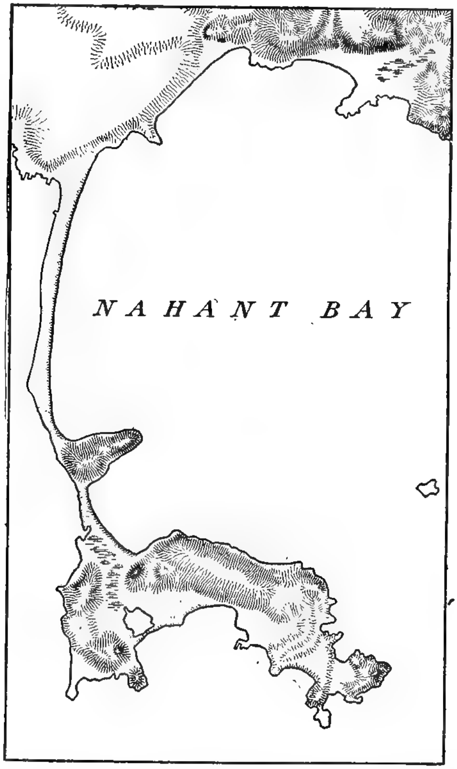


FIG. 71. — Former island of Big Nahant tied to Little Nahant, and the latter to the mainland by single tombolo.

in the lagoon at Venice, which are visible only at low water, are called "tomboli" in the Venetian dialect; and Prof. F. C. Ewart of Colgate University states that Petrocchi gives as one

* "'Tomboli' si chiamano comunemente per similitudine que' monticelli di rena, che il mare forma a guisa d'argini sulla spiaggia, altrimenti *Cotóni* = *Costoni* da *Costa:* per es, 'il mare tempestoso per traversia rade il fondo e riporta al lido quella rena, e forma i tumoli o i tomboli, e fa de' ridossi o interramenti alla bocca d'Arno. Essi tomboli sono la medesima cosa che le famose Dune degli Olandesi e Franzesi'. (Targioni, *Viaggi.*)"

PLATE XLII.

Tombolo connecting the former island of Marblehead with the Massachusetts mainland.

meaning of the word: "a small bank of sand thrown up by the sea." On maps of the Instituto Geografico Militare of Italy such names as "Tombolo della Giannella" and "Tombolo di Feniglia" are printed along bars connecting islands with the mainland. The fact that the singular form "tombolo" is used, rather than the plural "tomboli" suggests that the term refers to the bar itself, and not to the series of dunes which may occur upon it.

For the reasons outlined above, and for the further reason that the term tombolo has been introduced into a number of English discussions of shorelines, and even into some reports published in foreign languages, it seems advisable to adopt Gulliver's usage, rather than to use the double term "connecting bar" (which might equally well apply to a bay bar connecting two headlands), or to invent a new term. A single short term is desirable for the form under discussion, and notwithstanding the lack of uniformity and precision in the Italian use of the term "tombolo," its adoption into the English language with the restricted meaning given to it by Gulliver best meets the needs of the case.

If the former island is connected with the mainland or with another island by a single, simple bar, we have a *single tombolo* (Fig. 70 and Plate XLII). On the Massachusetts coast Big Nahant is tied to Little Nahant, and Little Nahant to the mainland by single tombolos in the construction of which onshore wave action on a shallowing bottom has probably played an important part (Fig. 71). A beautiful example of closely similar form is furnished by Duxbury Beach and Saquish Neck near Plymouth Harbor on the same coast (Fig. 72). Islands close to the mainland, or of comparatively large extent alongshore, may be connected with the mainland by a *double tombolo* or even a *triple tombolo.* Monte Argentario (Fig. 73) is tied to the west coast of Italy by a double tombolo, and a third uncompleted bar shows that the connection just escaped being a triple tombolo. Where two embankments extending backward from an island or outward from the mainland unite to form a single ridge before the connection is completed, we have a *Y-tombolo,* the type example of which is Morro del Puerto Santo (Fig. 74) on the Venezuelan coast[23]. *Complex tombolos* result when several islands are united with each other and with the mainland

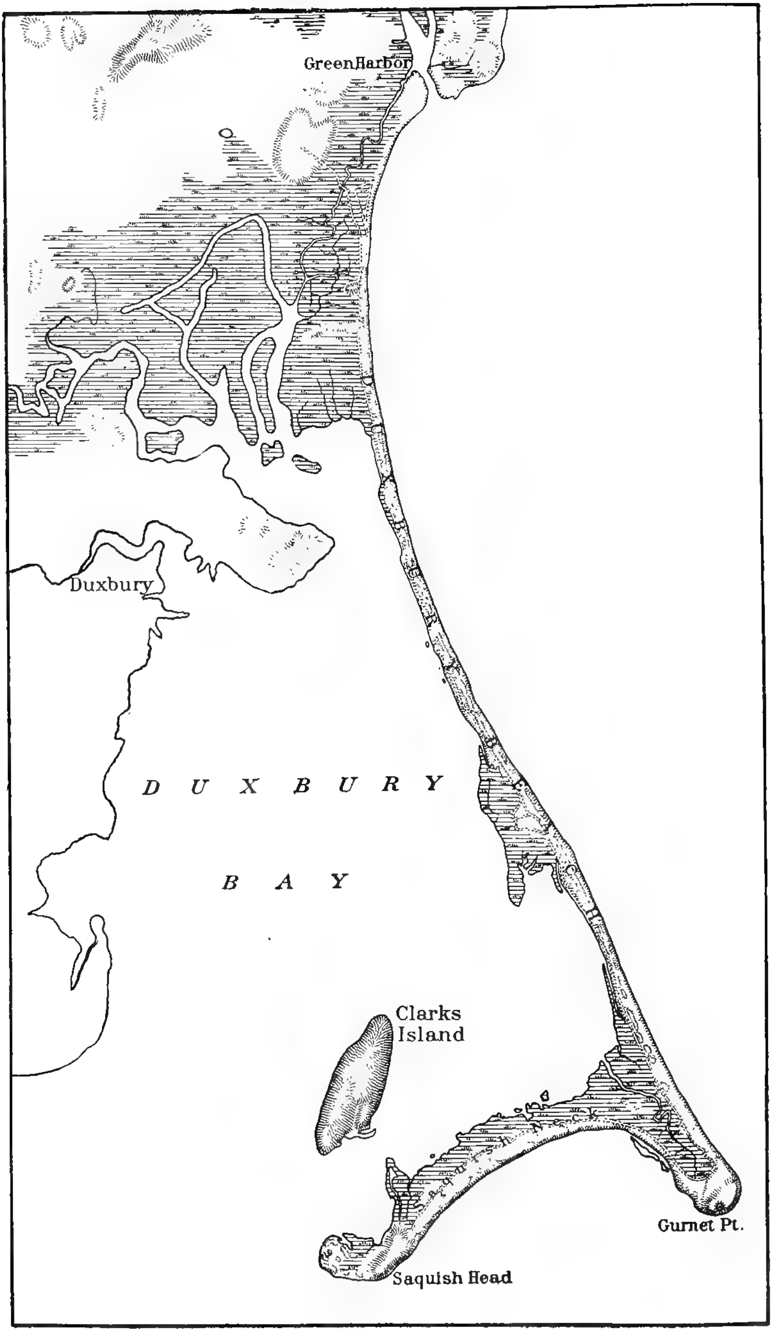


FIG. 72. — Duxbury and Saquish Neck tombolos uniting former islands with the mainland of Massachusetts near Plymouth.

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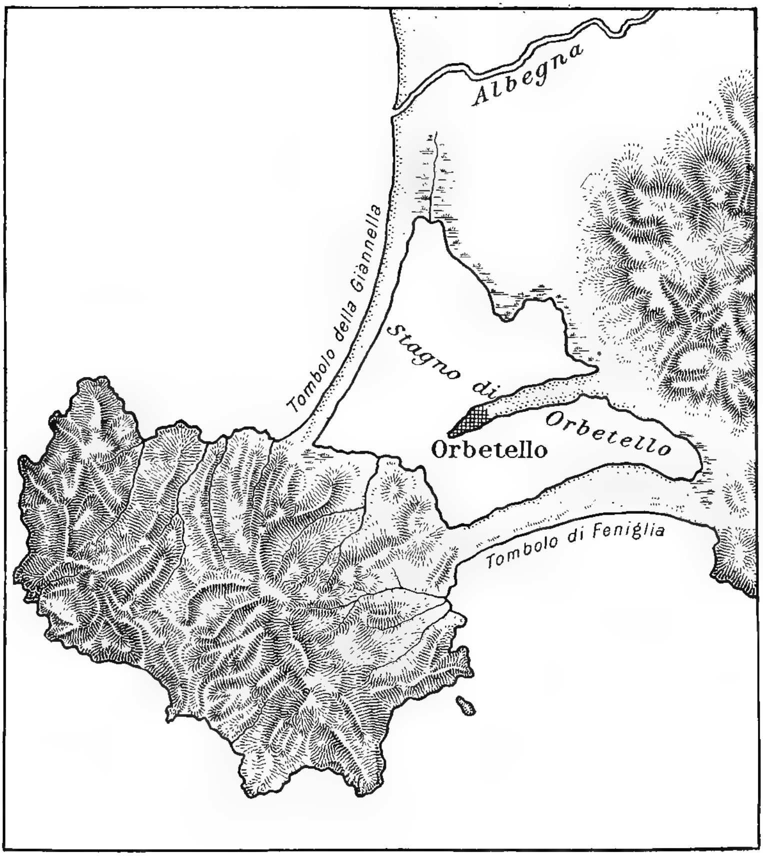

FIG. 73. — Monte Argentario, Italy, tied to the mainland by a double tombolo.

by a complicated series of bars. Nantasket Beach (Figs. 75 and 76) on the Massachusetts coast is an excellent example of this form, in which the prograding of some of the bars has produced a series of beach ridges extending over a breadth of half a mile. The complicated history of this remarkable tombolo has been fully discussed by Johnson and Reed[24]. It should be noted that the term tombolo refers to the connecting bar itself, and not to the former island as has been assumed by several, including Hobbs[25], who employs the spelling " tombola " and assigns to it a Spanish origin. This last was presumably due to an oversight, as I cannot find authority for a Spanish form of the word.

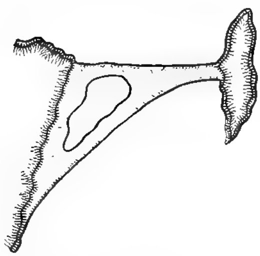

Fig. 74. — Morro del Puerto Santo, Venezuela, a Y-tombolo

Cuspate Bars. — It occasionally happens that a spit which has advanced some distance into open water, recurves (Fig. 77) until it again unites with the shore at its distal end, thus producing a bar which is more or less cuspate according as the seaward angle is fairly sharp or broadly rounded. An unusually sharp angle may result if a secondary spit trails abruptly back toward the shore from the point of a primary spit. In other cases two spits may grow out from the shore toward each other, and finally unite to form a bar of sharply cuspate form. This often happens on the leeward side of an island, when it represents the early stage of a Y-tombolo. Sometimes the presence of a shallow some distance out from the main shore will cause the development of a bar of similar form whose apex is at the shallow. Essentially identical in shape and origin is the bar which results when an island connected with the mainland by a double tombolo is consumed by the waves, leaving a V-shaped bar with the point of the V near the site of the former island. In both of the last two cases it is the obstruction in front of the main shore which determines the form and location of the resulting bar.

Certain features are common to bars developed in the manner above described. All of them are more or less cuspate in form; all enclose a lagoon or swampy area; and probably all have been produced by the combined action of waves and currents.

It is frequently impossible to determine the precise manner in which a given bar originated and developed. For this reason it seems wisest, as in the case of bay bars, to group the similar forms under a single name, recognizing the fact that different examples may have originated in different ways. The name V-bar has been applied to some of these forms; but because of their relation to cuspate forelands, described below, we will employ the term *cuspate bar* (Fig. 50, *cb*).

Were the compound spit (Fig. 55) which protects Toronto Harbor to unite with the shore at its distal end, as Fleming[26] considered a future possibility, we would have a *compound cuspate bar*. Caseys Point (Fig. 78) and Gaspee Point (Fig. 79), Rhode Island, representing what Gulliver[27] calls the V-bar stage and lagoon-marsh stage of cuspate forelands, are good examples of simple cuspate bars which were probably developed from spits growing seaward toward each other, or from primary spits growing seaward and secondary spits extending from their distal points backward to the shore. At the southern end of Revere Beach near Boston is a cuspate bar produced by the removal of an island which was connected with the mainland by a double tombolo (Fig. 80). The shores of Port Discovery on the Washington coast exhibit a beautiful series of cuspate bars in all stages of formation (Fig. 81).

Cuspate Forelands. — In none of the shore forms thus far considered has there been any extensive forward building of the main shore into the water. Beaches, spits, bay bars, tombolos, and cuspate bars are either comparatively narrow, or, as in the case of some broad spits and tombolos, are connected with the land by narrow embankments. We must now consider a group of forms in which the shoreline is systematically prograded by wave and current action, and an appreciable area of more or less continuous dry land added to that previously existing. The new land is sometimes called a beach plain; or, following Gilbert, a wave-built terrace. The latter term is more appropriate for the examples found on the elevated shorelines studied by Gilbert than for those on modern shores where the terrace effect is not evident because the top surface alone appears above water. We will follow Gulliver's suggestive terminology and speak of these features as *forelands*. They may have a variety of forms, but where most typically developed

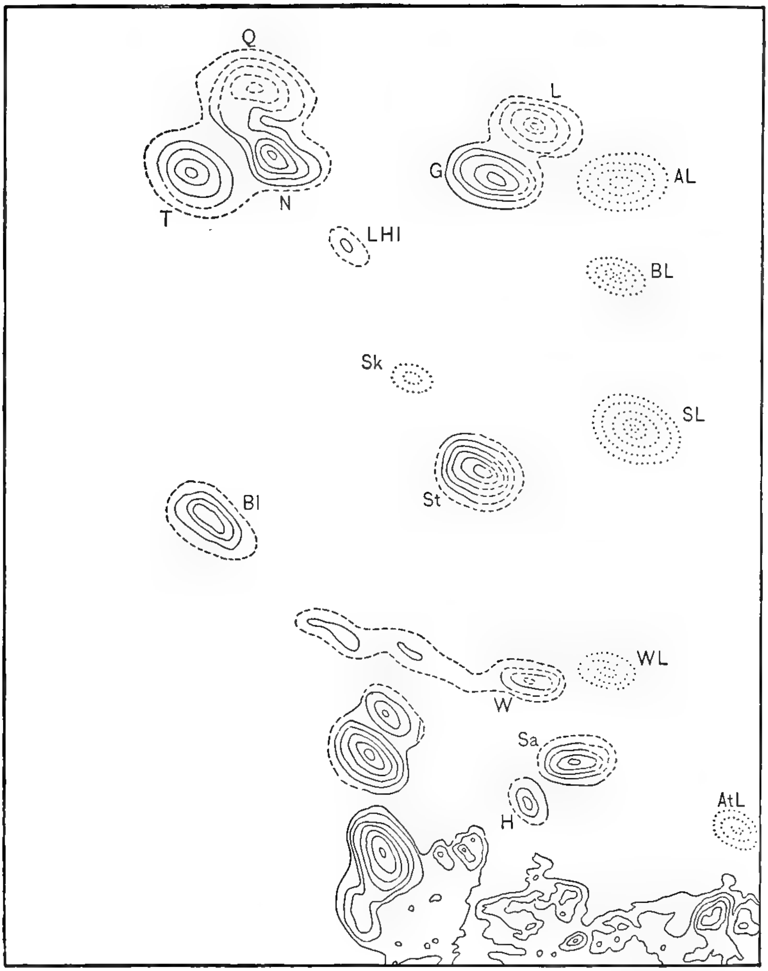

FIG. 75. — Former islands, many of which were wholly or completely eroded by wave action, and the resulting débris used by the waves to build a complex tombolo tying the remaining islands to the mainland. Dotted contours show islands wholly destroyed, broken contours the eroded portions of islands but partially destroyed. (Johnson and Reed.)

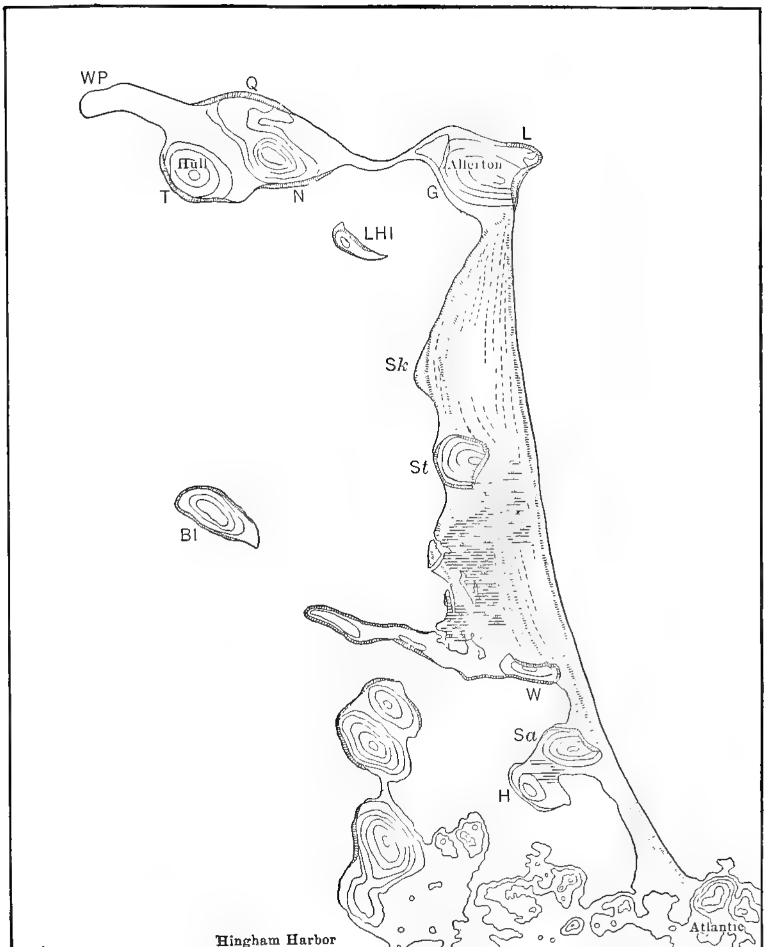

FIG. 76. — Nantasket Beach, Massachusetts, the complex tombolo formed by wave erosion of the islands shown in Fig. 75, with deposition of the débris to give connecting beaches uniting the remaining islands with each other and with the mainland at the south. (Johnson and Reed.)

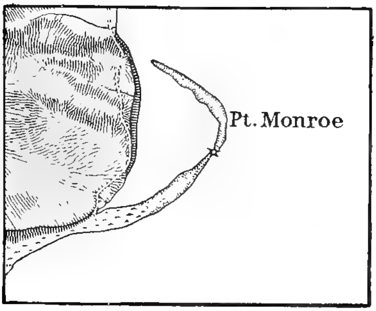

Fig. 77. A strongly recurved spit on the Washington coast, about to become a cuspate bar.

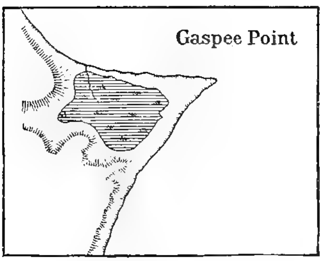

Fig. 79. — Cuspate bar showing enclosed marsh near Providence, Rhode Island.

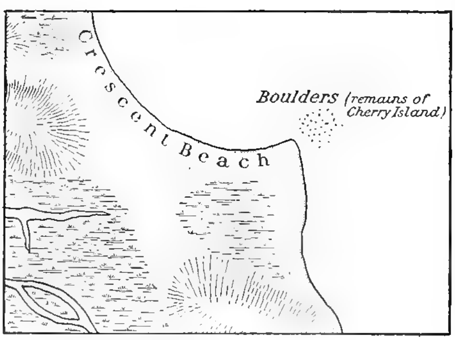

Fig. 80. — Cuspate bar originally built as a tombolo tying to the mainland an island since removed by wave erosion.

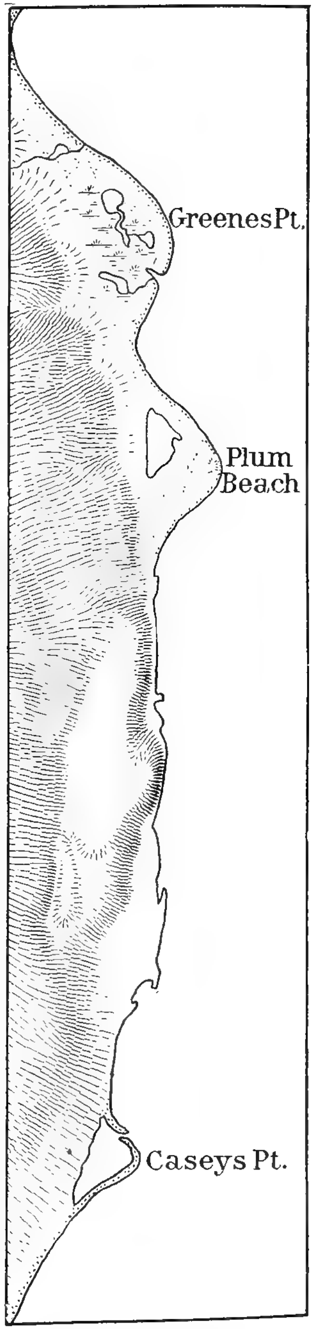

Fig. 78. – Cuspate bars on the Narragansett Bay shore.

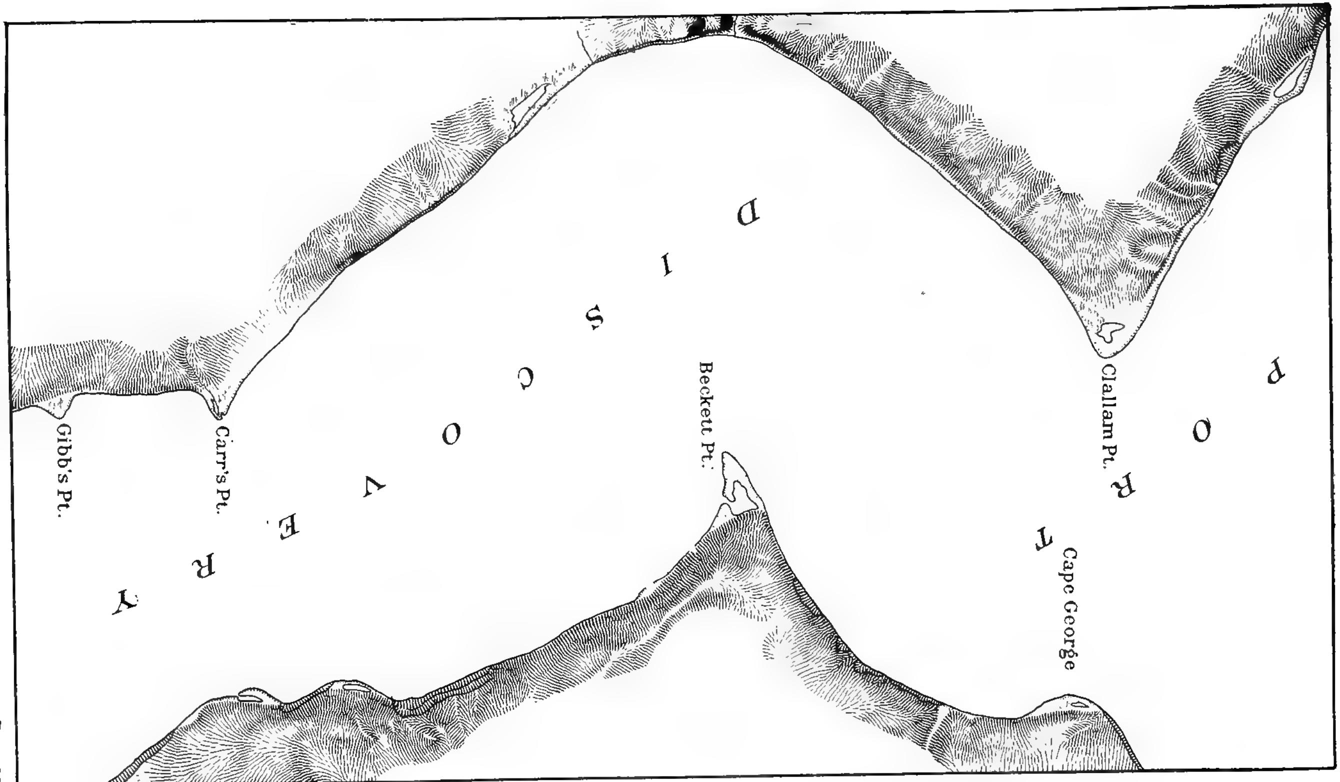

FIG. 81. — Cuspate bars on the shores of Port Discovery, Washington.

are more or less triangular in shape with the apex of the triangle pointing out into the water (Fig. 82); they are then called *cuspate forelands*[28]. A change in the outline of the shore or in the configuration of the sea-bottom often occasions their development, while in other cases no assignable cause is apparent.

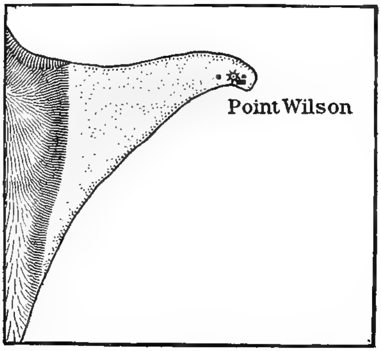

Fig. 82. — Cuspate foreland near Port Townsend, Washington.

There is no sharp dividing line between a compound cuspate bar in which the successive embankments are closely spaced, and a cuspate foreland in which the different beach ridges are widely enough separated to enclose strips of lagoon or marsh. Transition forms between the two types

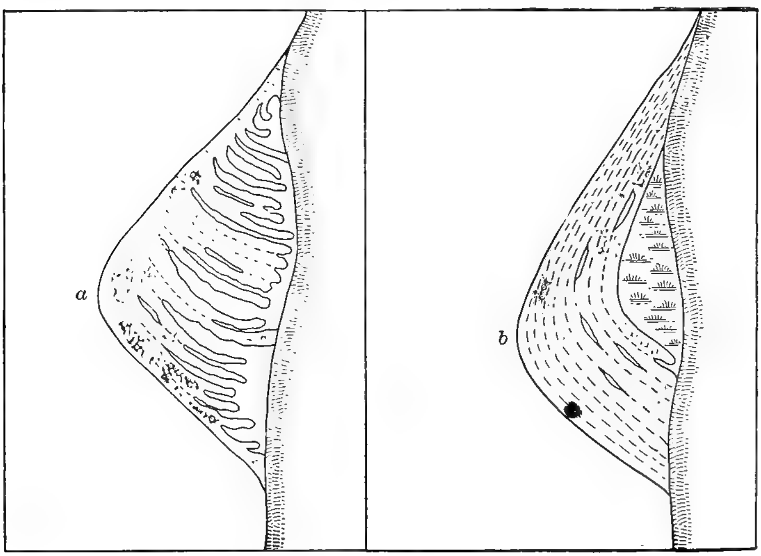

Fig. 83. — Types of cuspate foreland bars.

exist, and might appropriately be termed *cuspate foreland bars* (Fig. 83, *a*). Another intermediate form, properly classed under

the same term, is produced when a typical cuspate bar enclosing a triangular lagoon or marsh (Fig. 83, *b*) is prograded by the addition of successive beach ridges upon its seaward side. If the lagoon or marsh strips are a minor feature, or if the initial triangular lagoon is very small as compared with the total surface of added land, then the forms are called simply cuspate forelands.

I have found it profitable to recognize three principal types of cuspate forelands. When the shore is aggraded on both sides, so that fairly symmetrical lines of growth (beach ridges and swales) run parallel with both shores of the cusp, we have a *simple cuspate foreland.* In one of its former positions Cape Canaveral seems to have been a fairly good example of this type (Fig. 84). Where erosion attacks one side of the cusp to such an extent that no ridges and swales remain parallel to that shore, but the shoreline obliquely truncates these lines of growth, a *truncated cuspate foreland* is produced. As the type example of this form we might cite the Darss foreland on the Baltic coast of Germany (Fig. 131), whose western shore abruptly truncates a magnificent series of ridges and swales. Occasionally a truncated cusp of this type is later prograded, giving ridges and swales parallel to the new shoreline; and the process of alternate retrograding and prograding may be repeated a number of times with constantly varying direction. The resulting forms will be designated as *complex cuspate forelands.* To this class belongs the present Cape Canaveral, on which several distinct series of ridges and swales have been successively truncated (Fig. 129). The Dungeness (Fig. 130) of southeastern England is moderately complex near its seaward point.

It should be noted that Cape Canaveral occurs on a shoreline of emergence. It is cited here because cuspate forelands occur on all classes of shores, and because it affords unusually good examples of two of the three types of cuspate forelands defined above.

Marsh Bars. — An interesting form, not generally recognized, is produced by marine erosion of the seaward edge of a marsh which was originally unprotected from the sea by any barrier of sand or gravel. Wave attack separates the vegetable matter of the marsh from the sand which is usually present in greater or less amount, and casts the sand upon the edge of the remaining marsh in the form of a narrow ridge. On the map such

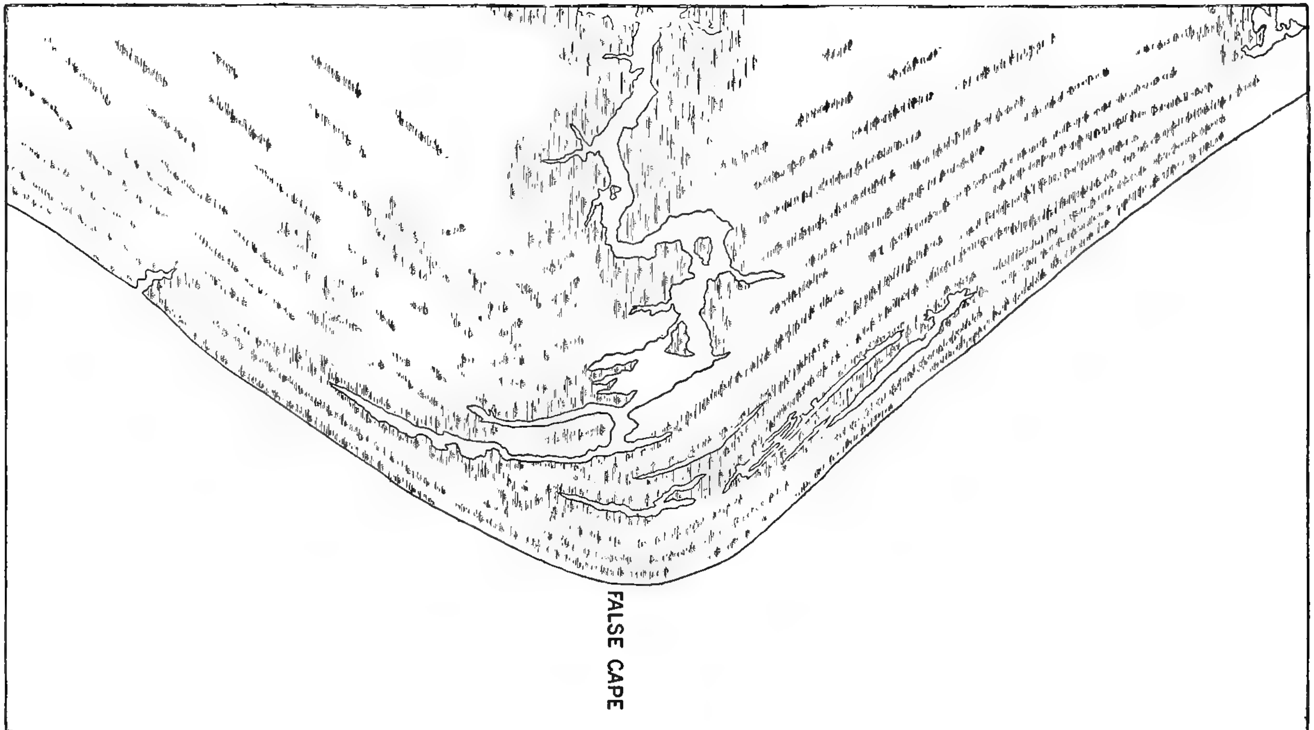

FIG. 84. — The former Cape Canaveral, now known as False Cape. Portions of the figure are restored, and all portions of the modern cape have been eliminated.

a ridge will look like a narrow offshore bar with a later formed marsh back of it. In reality the marsh is the older, and the ridge is quite unlike an offshore bar in origin. The smaller size, lack of continuity, and the irregular pattern of these *marsh bars* will generally distinguish them from true offshore bars. They are usually found bordering unexposed shores where unprotected marsh deposits could persist for a long time, suffering only gradual removal by small sized waves. Along the

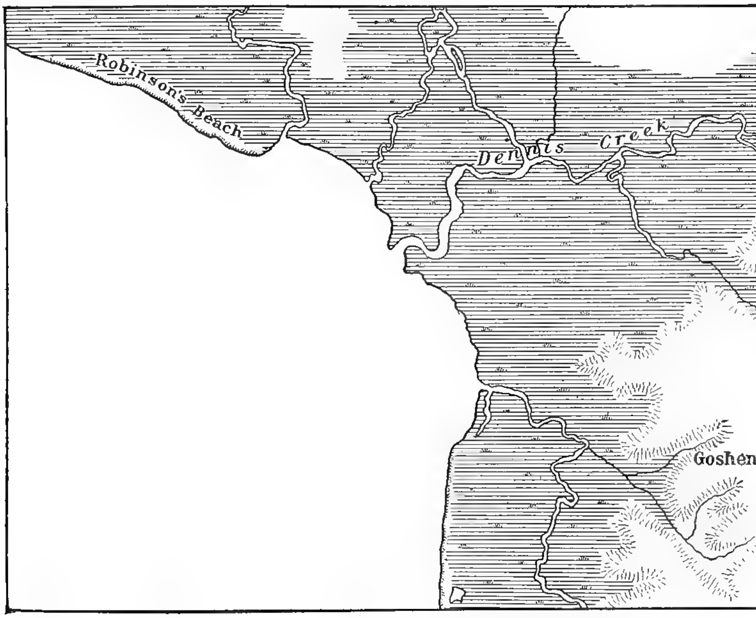

Fig. 85. — Marsh bars on the Delaware Bay shore.

Delaware Bay shores of New Jersey marsh bars are numerous, Robinson's Beach near the mouth of Dennis Creek being a good example (Fig. 85). The fact that some foreign débris may be brought to such a bar by waves and currents is immaterial, the essential point being that the marsh is older than the bar, and has never been bordered by a true offshore bar formed by wave action on the sea-bottom.

Flying Bars. — After a spit or looped bar has grown backward from an island, it sometimes happens that the island itself is entirely removed by wave attack before the spit or bar is de-

stroyed. We then have a *flying bar*, isolated in the open water. Gulliver[29], who originated the term, suggested that Sable Island, an isolated bar of unconsolidated sand off the coast of Nova Scotia, may be a flying bar left in its exposed position by the consumption of a former island to which it was attached.

Bay Deltas. — Streams entering the heads of drowned valleys will deposit sediment to form deltas, providing their currents transport more débris or débris of larger size than the marine currents in the bay can remove. The deltas normally advance from the heads of the bays toward the bay mouths, and may be designated by the term *bay deltas* (Fig. 50, *bd*). They often extinguish the lagoons left back of bay bars, or completely fill open bays, thus assisting the shore processes in their efforts to simplify the shoreline. It should be remembered, however, that they are the products of normal stream action, and are deposited in spite of marine processes, rather than because of them. For this reason it is a mistake to treat them as one of the forms resulting from the normal tendency of marine forces to simplify ragged coast. We must rather regard them as extraneous features whose effect in straightening the shoreline is wholly incidental and accidental, and quite independent of the processes by which waves and currents work toward the same result.

Stages of Development of Shore Details. — It may have been observed that in the preceding discussions of beaches, spits, bars, tombolos, forelands, and deltas, no account has been taken of successive stages of development of these forms. They have been described as forms especially characteristic of a young shoreline of submergence, but young, mature, and old stages of recurved spits, bay bars, and all the other forms mentioned, have not been recognized. The omission was intentional, and is due to the writer's doubt of the wisdom of attempting to classify the details of shore forms into definite stages of development. Inasmuch as this doubt has not been shared by all students of shoreline physiography, it is desirable that the grounds for its existence be made plain.

The greatest value of recognizing sequential stages of landform evolution is the aid thus given to a clear comprehension of the shape and significance of the forms in question. Assuredly, the introduction of the evolutionary idea into the study of river valleys, coastal plains, mountains, and other major land-

forms has shed a flood of light upon their present shapes and their past and future histories. We have seen that the shore profile cannot be fully understood except in the light of its successive and orderly stages of development; and the same is true of the outline of the shore as a whole. On the other hand, it may well be doubted whether it is profitable to push the developmental idea so far as to apply it in the explanation of all the detailed forms which are merely incidents in the evolutionary history of some major topographic unit. The term "young river" is full of significance for the student of land-forms; but I doubt whether anyone will profit from an attempt to recognize young, mature, and old stages of sandbars, which may occur in any or all of the different stages of river development. Similarly, I find unlimited value in the recognition of young, mature, and old stages of shorelines; but am not convinced that there is profit in the effort to classify all the details of a young shoreline, for example, into three or more special stages of development. Unless it shall appear that the understanding of shore forms is materially aided by such attempted classification, we may better restrict the application of terms indicative of developmental stages to the shoreline as a whole, rather than extend their use to each of its many parts.

A further reason for not recognizing definite successive stages in the development of spits, bars, forelands, etc., is the difficulty of determining any regular and orderly succession of features which will be common to all forms of a given class, and which are genetically related to true shoreline processes. The best attempt to classify shore details into stages of development is that made by Gulliver; but the results of that attempt are not altogether satisfactory. Thus the youth of a tombolo is assumed to be represented by one or two cuspate forelands projecting from mainland toward island, or island toward mainland, or both, even though the intervening channel may be so deep that further growth of the forelands is impossible.[30] On this basis, a tombolo which had been entirely completed, and then broken through by storm waves, would be called "young." A completed tombolo is said to be in "adolescence," or, if the island happens to be nearly or quite eroded away, "late adolescence;" while "the mature stage of island-tying is where the islands and their connecting tombolos are completely consumed

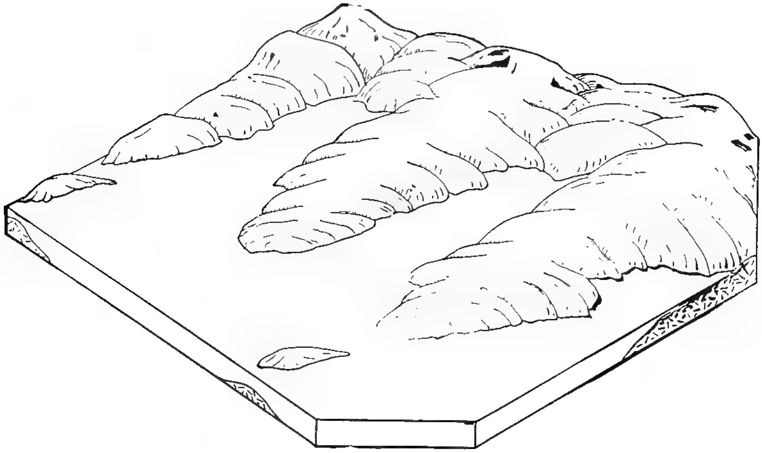

Fig. 45.

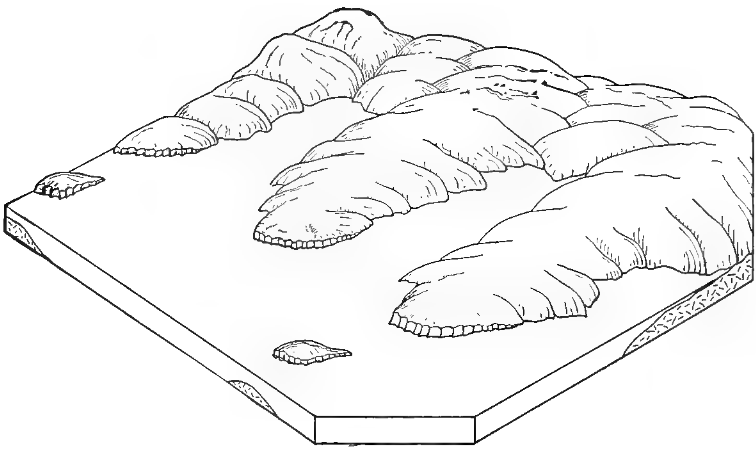

Fig. 47.

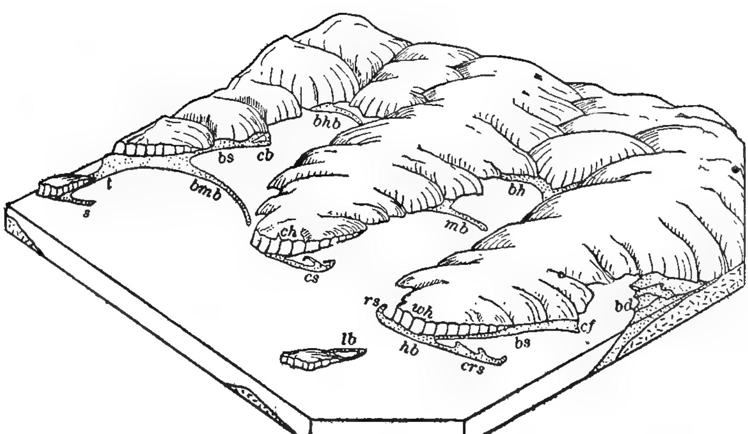

Fig. 50.

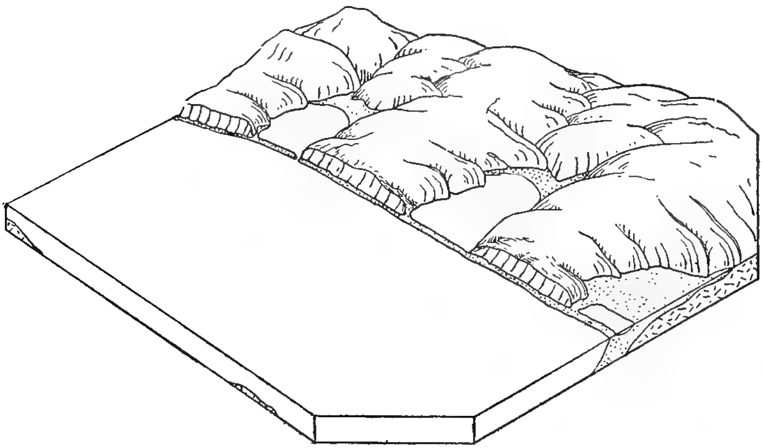

Fig. 87.

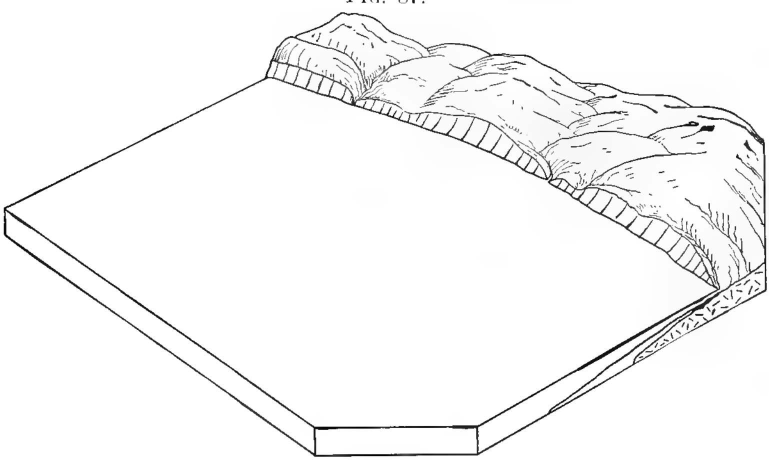

Fig. 88.

Comparison of text figures to facilitate correlation of successive stages in the development of a shoreline of submergence.

Fig. 45. — Initial stage.
Fig. 47. — Early youth.
Fig. 50. — Youth.
Fig. 87. — Submaturity.
Fig. 88. — Maturity.

bd, bay delta; *bh*, bayhead beach; *bhb*, bayhead bar; *bmb*, baymouth bar; *bs*, bayside beach; *cb*, cuspate bar; *cf*, cuspate foreland; *ch*, cliffed headland; *crs*, compound recurved spit; *cs*, complex spit; *hb*, headland beach; *lb*, looped bar; *mb*, midbay bar; *rs*, recurved spit; *s*, spit; *t*, tombolo; *wh*, winged headland.

by the sea[31]." Surely the developmental idea is forced beyond the limits of its usefulness when the complete annihilation of a given form is called its "maturity." One reason for the unsatisfactory character of this classification is that it represents an attempt to harmonize the stages of tombolo formation with the stages of shoreline development, an attempt which must always end in failure for the reason that the isolated, detailed forms of an irregular shoreline, even if they develop systematically, cannot develop synchronously with the shoreline as a whole. Tombolos may be made and destroyed while the shoreline is still in its youth.

Gulliver's attempt to classify bay bars according to stages of development is equally unsatisfactory. The basis of classification was made the extent to which the bay was filled by a stream delta or other deposits. Since these deposits are quite independent of the bar itself, and are found abundantly in bays which have no bars, they can scarcely be accepted as a proper basis for the classification of bars into young, adolescent, and mature examples. If bay bars have any orderly sequence of forms characteristic of different stages of their development, they must be classified, if at all, on the basis of those forms, and not on the relative size of wholly extraneous features, such as river deltas, which may happen to lie back of them.

In discussing cuspate forelands Gulliver drops the terms youth, adolescence, and maturity, for reasons which are not clear, and speaks of "three stages of progressive development,—the V-bar stage, the lagoon-marsh stage, and the filled stage." He recognizes that the first two stages are not represented in the history of those forelands which build out continuously from the mainland. Bay deltas are classified as young, adolescent, or mature according to the extent to which they fill the bay into which they happen to be built. Of three deltas identical in size, shape, and composition, but built into three bays of increasing length measured from head to mouth, one would be called mature, another adolescent, and the third young. Ordinary deltas are classified, not according to stages of development, but according to form as determined by the ratio of activity between river and marine currents, because it was not found practicable to discover laws of progressive delta development when the deltas did not occur in bays. This fact must lead us

to question the value of classifying into definite stages of development those deltas which happen to be located in bays; especially when such classification is based, not upon real differences in the characteristics of bay deltas at different periods of their formation, but upon the non-significant ratio of delta size to size of bay. Gulliver makes no attempt to divide spits into stages of development[32].

Enough has been said to show the difficulty of classifying the details of shore forms into progressive stages of systematic development. It is clear that, at least in the present state of our knowledge, such classification is neither profitable nor feasible. This conclusion is perfectly compatible with the belief that shore profiles and shore outlines pass through perfectly definite stages of development, the proper recognition of which is essential to a full understanding of shore forms. Gulliver rendered a valuable service to physiography by applying the principles of landform evolution to the study of shorelines on a scale never before attempted. That he may possibly have carried the attempt too far does not affect the fundamental importance of his thesis.

Relative Importance of Different Marine Forces in the Formation of Bars, Forelands, Etc. — Throughout the discussions of beaches, spits, bars, tombolos, and forelands which have occupied our attention on preceding pages, no special consideration was given to the marine forces which produced those forms. It was stated that waves or currents operated in certain ways, but ordinarily neither the methods of wave action nor the kinds of currents were discussed. In previous chapters we have analyzed the behavior of waves and currents of different types at some length; but it remains to answer the important question as to which of these agencies are primarily responsible for the detailed forms found on a young shoreline of submergence.

Gulliver[33] recognizes three marine agents: waves, tides, and currents. A careful reading of his essay on "Shoreline Topography" shows that under "waves" he does not clearly recognize the highly important wave currents, but only the destructive effects of wave impact; by "tides" he means tidal currents; and under "currents" he refers to planetary currents and local wind currents. He is "inclined to attribute the attack of the sea largely to the waves, and its transporting action largely

to the tides and currents;" and throughout the discussion of individual shore forms he adheres to this idea of transportation largely by tidal, planetary, and wind currents. There are, it is true, isolated statements which taken alone seem to indicate a fuller recognition of wave-current action; but the treatment as a whole practically excludes this important process. Thus in discussing the origin of cuspate forelands in estuaries Gulliver shows that planetary currents cannot operate in such localities, and that wind currents are so weak as to be overpowered by the tides; he therefore concludes that tidal currents must be responsible for the forms in question. Wave currents and the associated "beach drifting" are not even referred to in this connection. The failure to recognize the very great efficiency of wave currents in moving shore débris is responsible for the idea, repeatedly expressed in Gulliver's essay[34], that important longshore transportation does not take place until more waste is supplied to the sea than can be deposited offshore. This might be true if, as Gulliver supposed, shore débris were dependent upon tidal, planetary, and wind currents for its transport; for not until the irregularities of sea-bottom and shore outline have been measurably smoothed out by local deposition, or the shore has reached its "adolescent stage" according to Gulliver, can these larger currents sweep uninterruptedly along the coast. Wave currents, however, will operate effectively on any shore which is fronted by a body of water sufficiently large for the generation of waves; and the most irregular shore will, even in its youthful stage, experience a very considerable amount of longshore beach drifting.

There can be no doubt that wave currents and the associated longshore beach drifting play a very important rôle in the formation of various types of beaches, spits, bars, tombolos, and forelands. Tarr[35] has shown that cuspate forelands, bay bars, tombolos, and spits are built by wave action in lakes and nearly tideless bays where tidal and other currents are either wholly inoperative or far too weak to move the material with which the forms have been constructed. Woodman[36] has presented convincing evidence that in the Bras d'Or Lakes of Cape Breton Island cuspate forelands, tombolos, bay bars, loop bars, and spits are formed by wave action without material aid from

tidal, wind, or other currents. Wilson's studies[37], on the shore forms of Lakes Erie and Ontario, and the Bay of Quinte lead to a similar conclusion. The tideless shores of the island of Rügen in the Baltic, as described by Philippson[38], exhibit numerous spits and bay bars composed of material transported almost exclusively by wave currents. A small cuspate foreland on the shore of Lake George is described by Comstock[39], as having been formed through the action of waves generated by passing steamboats.

Even where tidal and other currents not related to wave action move with high velocity in the offshore zone, the waters near the shore, where the forms in question are built, commonly have a movement too feeble to transport the gravel and cobblestones of which many forelands and embankments are composed. On the other hand, wave currents near the shore are exceedingly powerful, and may easily be observed to drive the coarsest débris along the coast with a rapidity which is sometimes surprising. Wheeler[40] repeatedly observed half bricks on a shingle beach carried 25 to 30 yards in from $1\frac{1}{2}$ to 2 hours, and quotes de Rance as authority for the drifting of encaustic tiles by a gale for a distance of "1 mile in two tides." Shaler reports the movement of pieces of brick by oblique wave action at the rate of more than half a mile per day. Wind currents in the shallow waters near the shore, and hydraulic currents generated by the combined action of waves and wind, while generally too feeble to move coarse débris without the aid of wave currents, frequently co-operate in a most effective manner with wave currents in causing a comparatively rapid and exceedingly important longshore transportation of both fine and coarse material. It is often feasible to demonstrate that the material of a given foreland or embankment is derived from a neighboring cliff, that beach drifting from the cliff toward the area of accumulation proceeds actively under the influence of waves, and that no other type of currents are known to exist which are of sufficient velocity to move the débris undergoing transport. It would seem logical to conclude that wave currents are mainly responsible for the production of the forms in question.

It has sometimes been held that the waves merely agitate the débris near the shore and by repeatedly raising it from the bottom make it possible for even weak tidal currents to effect a

longshore transportation of coarse material. There can be no doubt that tidal and other currents often co-operate with wave currents to effect the distribution of shore débris; but it should be remembered that wave currents are independently capable of moving the coarsest material along the shore for indefinite distances. Gravel and cobblestones would be carried along a coast by wave currents and built into various types of forelands and embankments, even were there no assistance from tidal and other currents; but these latter currents would in general be powerless to move such coarse débris in the immediate vicinity of the shore unless aided by waves; a fact fully appreciated by Gilbert[41]. I find it impossible, however, to accept Gilbert's further conclusions that "the transporting effect of waves alone is so slight that only a gentle current in the opposite direction is necessary to counteract it," and "the concurrence of waves and currents is so general a phenomenon, and the ability of waves alone is so small, that the latter may be disregarded[42]."

One must, however, fully recognize the possibility that tidal and other currents may be primarily responsible for the location and development of some forelands and embankments. For the production of these forms it is only necessary that shore débris shall be transported to a certain locality and there deposited. It is immaterial what type of current accomplishes the transportation. If tidal currents, or eddy currents, or currents of any other type have the proper direction and strength to accomplish the observed results, their possible importance must not be overlooked simply because wave currents are known to have produced similar results elsewhere. The possibility that certain sandy cuspate forelands, spits, etc., are primarily the product of currents unrelated to wave action should especially be kept in mind. It may be difficult, or even impossible, to determine the relative importance of wave currents and other currents in such cases; but if the determination is at all possible, it can safely be made only by one who studies the individual examples in the field with an open mind, and who is fully convinced of the ability of wave currents, as well as other more generally recognized currents, to produce such forms. Abbe[43] reports a case in which an eddy current generated by the ebbing tide seemed to him to be responsible for the development of a cusp on the sandy shores of Sassafras River in Maryland. As

regards my own studies, I can say that I have found many forelands and embankments which seemed to me demonstrably due, principally if not wholly, to wave currents; but none which seemed undoubtedly the product of other types of currents. I am therefore inclined to believe that wave currents have played the most important part in the construction of all the sandy forelands and embankments of our coasts.

Davis[44] refers briefly to an interesting cuspate bar on the south shore of Lake Balaton in Austria-Hungary, the posi-

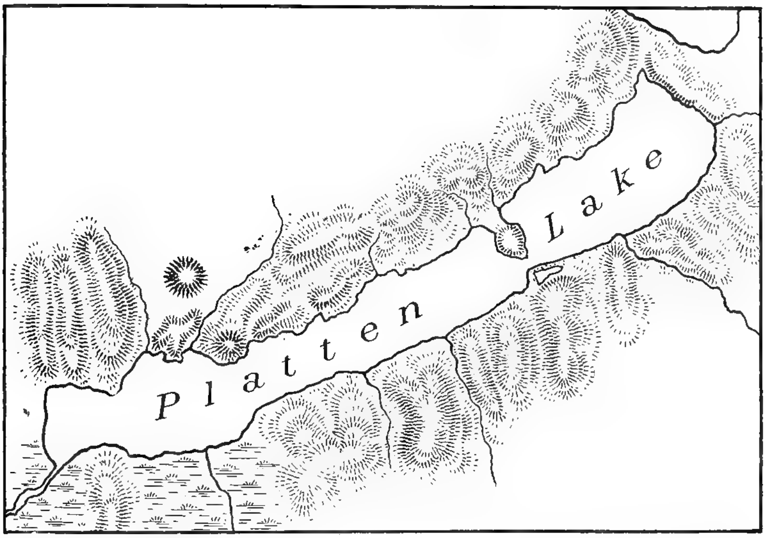

FIG. 86. — Lake Balaton (Platten Lake) showing position of cuspate bar.

tion of which he regards as evidence that currents rather than waves control the development of such forms. On the basis of map study he concludes that a promontory which projects far into the lake from the north shore probably occasioned the development of two circling currents, and that in the quiet water between the two whirls, near the south shore, the cuspate bar was built. An outline map of the lake, showing the position of the bar, is represented in Figure 86. When due account is taken of the relation of fetch of open water to wave development, it will be seen that westerly winds must drive large waves eastward along the south shore of the western arm of the lake

as far as a point opposite the projecting promontory; but that beyond this point wave action from the west will be weakened because of the sheltering effect of the promontory and the resulting short stretch of open water across which west winds can blow. Northerly and northeasterly winds will drive fairly large waves against the south shore of the eastern arm of the lake, but just west of a point south of the promontory there will be a rapid decrease in the comparative effectiveness of waves from this direction. As a result of these conditions we should expect beach drifting along the south shore of the lake to be toward a point opposite the promontory for a considerable distance on either side of that point. It is not necessary, therefore, to assume the existence in Balaton Lake of two rotary currents of sufficient velocity near the shore to transport the débris which composes the cuspate bar, for a bar opposite the promontory is a perfectly normal and expectable result of wave action alone. It may even be shown that wind waves from a single direction are alone competent to build a cuspate bar at the point in question.

The frequent appeal to a pair of hypothetical circling currents or eddies with a triangular space of comparatively dead water between the shore and the point of tangency of the eddies, in order to explain the development of cuspate bars and forelands, has long seemed to the writer unnecessary, and insufficiently justified by evidence of critical value. In many cases there is ample evidence that currents of some type effect the longshore movement of débris either toward or away from the point of the cusp; but in most cases the only basis for the supposed pair of circling currents is the assumption that they are required in order to explain the presence of a foreland of cuspate form.

Both the development of cuspate bars and forelands, and the longshore movement of débris causing offsets, overlaps, and stream deflections, may usually be explained as the normal product of longshore beach drifting, assisted by the wind currents and hydraulic currents which ordinarily accompany that process. Thus the Darss foreland (Fig. 131) has been built with débris drifted eastward by the action of waves generated under westerly winds, the beach drifting no doubt having been supplemented by water forced eastward by the friction of the wind

on the sea surface, and by eastward moving hydraulic currents which would attempt to remove the water piled against the coast by wind and waves. As the point of the Darss advanced northward it gradually sheltered the waters to the eastward from westerly winds, and gave the waves generated by easterly winds, formerly overpowered by the dominant action from the west, an opportunity to determine the movement of shore débris. Consequently beach drifting from east to west has apparently prevailed east of the northward projecting point of the Darss in recent times, and has doubtless caused the westward deflection of the Prerow River. In a similar manner the Carolina capes, regarded by Gulliver[45] and Davis[46] as having been built between pairs of circling currents, may be explained as the expectable result of normal wave action. In neither case, nor in any other known to the writer, does it seem necessary to assume the existence of pairs of rotary currents, the evidence for which is either inconclusive or wholly lacking.

Mature Stage. — We have inquired at some length into the series of forms which characterize the young shoreline of submergence, and have found that the unorganized condition of current action along such a shore combines with the initial irregularities of the submerged land area to produce an almost endless variety of interesting shore features. In striking contrast with the complexity and variety of youth is the simplicity of the mature shoreline of submergence (Fig. 88). Let us trace briefly the steps by which that simplicity is attained. During its initial stage a shoreline of submergence is wholly unadjusted to the waves and currents which operate upon it. Waves break irregularly upon the uneven bottom and against the complicated shoreline; currents are split up and deflected in every conceivable direction, and any branch current may find itself flowing swiftly against some headland or over some shallow at one moment, and dropping its load a moment later when its velocity is checked upon passing into deep water opposite some bay. This unadjusted condition continues in constantly diminishing degree throughout the youth of the shoreline of submergence and is characteristic of that stage of its development, just as an irregular longitudinal profile is characteristic of the young stage of stream development. But the removal of outlying islands, the cutting back of projecting

headlands, and the building of bars across the mouths of bays, gradually simplify the outer shoreline and permit longshore currents to move through greater distances unimpeded by projecting land masses. At the same time wave action has established the shore profile of equilibrium on the seaward side of the bars and is working toward the same end on the cliffed headlands. Thus the shoreline progresses toward maturity. When the headlands are partially cut back and many of the intervening bays are nearly or quite closed by bars, so that longshore currents may move through considerable distances before en-

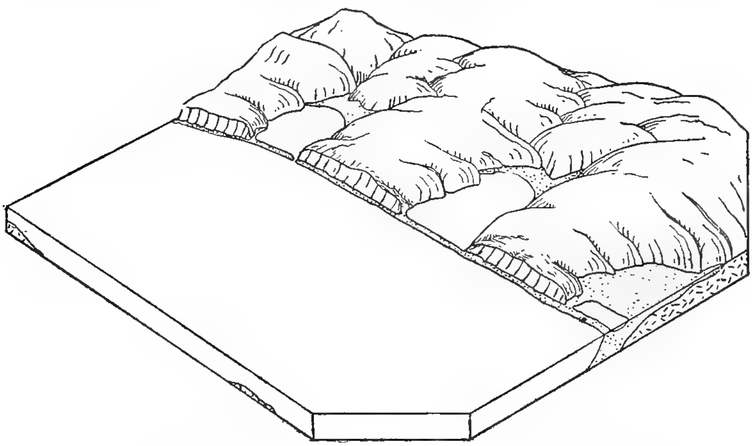

FIG. 87. — Shoreline of submergence, submature stage.

countering obstructions, the shoreline may be said to have reached late youth or submaturity (Fig. 87).

Still later the headlands will be so far cut back and bay-mouth bars will be so uniformly present opposite the original re-entrant angles of the coast, that a very simply curved or nearly straight shoreline will permit longshore currents to transport débris for indefinite distances without hindrance. The shoreline is now nicely adjusted to the forces operating upon it; the beach profile of equilibrium is fully established both on the seaward side of the bars and at the bases of the retreating cliffs; and while the cliff profile may still be too steep to permit one to call the entire shore profile mature, the shore outline is such that débris moves with the longshore currents as systematically as

sediment is moved seaward on the nicely adjusted slope of a graded river. In short, the shoreline itself has reached a graded condition; and this condition has been attained by an orderly process of cutting back the headlands and bridging the bays with the resulting débris, just as the grading of the river is accomplished by cutting down projecting rock masses and filling depressions with the erosion products. The establishment of the graded condition marks the entrance of either river or shoreline into the early mature stage of its development.

Full maturity (Fig. 88) is attained only when the shoreline has been pushed inland beyond the bay heads, and lies against the

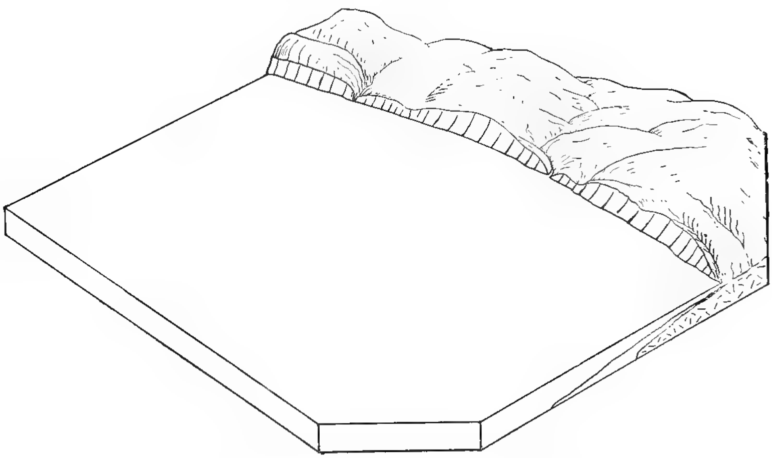

FIG. 88. — Shoreline of submergence, mature stage.

original mainland throughout all its course. By this time the numerous islands and prominent headlands of youth have been obliterated, the great variety of spits, bars, tombolos, and forelands have disappeared, bay deltas and marsh deposits have been consumed by the advancing waves, and there remains only a comparatively narrow beach at the base of an almost continuous marine cliff which borders a shoreline of very simple curvature. Monotony rather than variety is the distinguishing feature of maturity, although the cliffs, often covered with vegetation, may be of such magnitude as to impart majestic grandeur to the coastal scenery.

Even in the maturity of a shoreline of submergence the ad-

PLATE XLIII.

Hanging valleys on the chalk coast of southeastern England, where waves cut back the shore faster than the streams can deepen their valleys.

vance of the waves may be so rapid that stream erosion cannot lower valley floors as fast as the shoreline is cut back. This is especially apt to be the case where streams are small and weak, and where the nature of the country rock permits much underground seepage and little surface erosion. Hanging valleys due to this cause are developed on a small scale along many coasts (Plate XLIII), but are especially striking on the mature coast of northwestern France (Plate XXI), where many valleys are left hanging in the face of the chalk cliffs because wave erosion

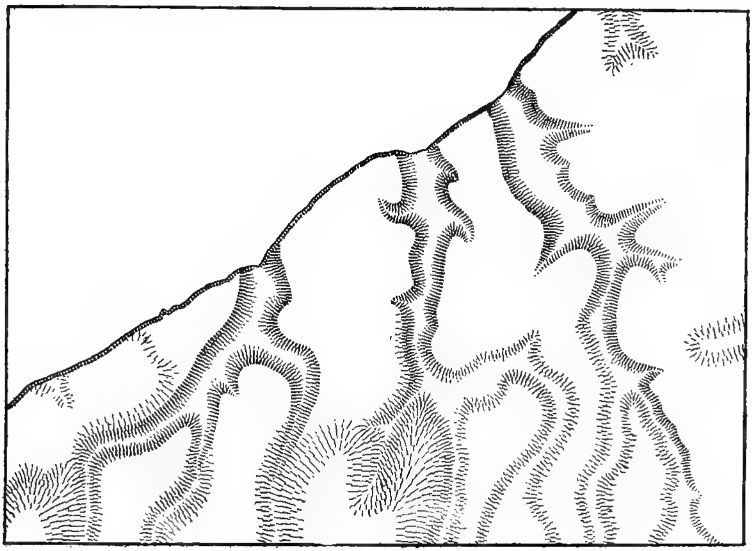

FIG. 89. — Valleuses on the northwest coast of France.

cuts the cliffs backward faster than the smaller streams can cut their valleys downward. These hanging valleys have been given a special name, "valleuses," by the French, and by their frequent convergence toward a point some distance out to sea (Fig. 89), they furnish an indication of the extensive wave erosion which has removed the main valley they once united to form. Only the larger streams have been able to reduce their gently sloping valley floors to sealevel as fast as the waves cut inland, and their valleys form the only interruptions in a line of cliff which extends for many miles in very simple curves. Where streams are unable to reduce a whole broad valley as rapidly as the shoreline is

worn back, we may find a narrow gorge cut in the bottom of the broad valley, the bottom of the gorge alone being reduced to an accordant junction with the sea.

Not all parts of a shoreline develop at the same rate. Along weak rock coasts maturity is attained more quickly than along coasts composed of more resistant material. Even after maturity is attained the shoreline on a broad belt of weak rocks will retreat more rapidly than adjacent sections, until the depth of the indentation has so weakened wave attack that a condition of equilibrium is attained. Thereafter all parts of the shoreline retreat at the same rate, the portion bordering the weak rock area keeping a constant distance in advance (farther inland). The initial more rapid retrogression of the shoreline on weak rocks depends primarily on two factors: in the first place the weak rocks yield more readily to wave attack; and in the second place, weak rock areas are normally worn down nearer to sealevel by subaërial agencies, with the result that waves and currents have to dispose of much less débris than they do where high cliffs shed vast quantities of waste upon a slowly retrograding beach. A mature coast should therefore show simple but distinct curves systematically related to rock structure.

Exposed shorelines develop more rapidly than do protected shorelines, a fact well illustrated by the more advanced stage of development reached on the Atlantic coast of the Maryland-Delaware coastal plain, as compared with the Chesapeake Bay coast of the same district. Other factors likewise retard or accelerate shoreline development, with the net result that a shoreline approaching maturity may consist of a series of more or less isolated stretches in a mature stage, separated by other stretches which are still submature or even young. As the waves cut farther into the land the mature sections increase in length, and finally unite when the entire shoreline has attained full maturity.

Old Stage. — The old age of a shoreline of submergence does not differ essentially from the old age of a shoreline of emergence. It will be more convenient, therefore, to postpone discussion of this stage of development until after the youth and maturity of shorelines of emergence have been considered.

RÉSUMÉ

In the present chapter we have passed in review those shore forms which characterize the different developmental stages of shorelines of submergence. It has been shown that by far the greatest variety of forms is associated with young shorelines of this class; while in the mature and old stages the forms are fewer in number, more simple in character, and more nearly like those in the corresponding stages of other classes of shorelines. Special consideration has been given to the question as to how far it is wise to attempt the classification of minor details of shore form into successive stages of development, and reasons presented in support of the opinion that these minor forms should not be so classified. The origin of the various shore forms, including cuspate bars and forelands, have been examined with some care in order to discover which marine forces are primarily responsible for their development; and the conclusion has been reached that beach drifting under the influence of wind-formed waves is more potent in their construction than are tidal and other currents. We are now prepared to turn our attention to shorelines of emergence, and to discuss the special forms characteristic of their successive developmental stages.

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CHAPTER VII

DEVELOPMENT OF THE SHORELINE (Continued)

B. SHORELINES OF EMERGENCE

Advance Summary. — The method of treatment followed in the preceding chapter is here applied to shorelines of emergence. Features characteristic of the youth, maturity and old age of shorelines of this class are described, and special emphasis placed upon those forms which for any reason merit extended consideration. Thus the origin of the offshore bar is quite fully discussed, and new evidence presented to test conflicting theories. The history of tidal inlets is traced in some detail, and in view of their behavior modifications of the current explanations of offshore bar development are suggested. It is shown that there exists a significant relationship between the positions of headlands to which some offshore bars are attached, the direction of longshore currents, the distribution of inlets, the width of lagoon and the extent of lagoon filling; and an explanation of this interesting relationship is offered. The effect of coastal subsidence and coastal elevation upon the history of the offshore bar and lagoon are discussed, and the fallacy of the theory that offshore bars are an evidence of coastal subsidence is exposed. Such an account is given of the changes to which offshore bars, tidal inlets and lagoons are commonly subject, as will, it is hoped, prove of value to the harbor and marine engineer as well as to the geographer and geologist.

Initial Stage. — When a sea-bottom or a lake-bottom emerges from beneath the water, either because of an uplift of the land or a sinking of the water surface, the new shoreline may be called a "shoreline of emergence." The essential characteristics of such a shoreline depend upon the fact that the bottoms of lakes and seas are not subjected to the river erosion which roughens land surfaces, but on the contrary are made even smoother by the continual deposition of matter brought into these quiet water bodies. If the plain of deposition emerges,

the water surface coming to rest against any portion of the nearly level plain surface, will give a straight or nearly straight shoreline.

In case the bottom of a sea or lake represents a rugged land area but recently depressed, and the emergence occurs before deposition has had an opportunity to bury the inequalities and produce a smooth subaqueous surface, then the emergence of the rugged bottom will give an irregular shoreline. Evidently the dominant features of this shoreline were determined by the submergence of the original hills and valleys, and not by the later partial emergence which was insufficient to change the type of the shoreline already existing. For purposes of classification and study such a shoreline must be grouped with shorelines of submergence, the partial emergence being of secondary importance only. Thus the coast of Maine is an excellent example of a shoreline of submergence, although a moderate amount of emergence succeeded the submergence which gave the shore its essential characters.

A subaqueous plain of deposition normally has a surface gently inclined away from the shoreline. After emergence, therefore, we should expect to find shallow water seaward from the new shoreline of emergence, the offshore slope being very gradual This is one of the essential characteristics of the initial stage of a shoreline of emergence; and since the shallowness of the water prevents the access of large waves to the shore, the early stages of development of such a shoreline are much affected by this feature.

During storms large waves break far out to sea, sometimes encountering water too shallow for their propagation several miles from the shoreline. Smaller waves reach the shore and begin their attack upon the land. A cliff is cut, which, because of its small size, is sometimes called a *nip* in the edge of the land. In the manner fully explained on previous pages the bench in front of this small cliff is gradually deepened and the cliff pushed inland, increasing in height as it is cut farther into the upward sloping coastal plain. In the meantime the large waves breaking farther seaward are cutting into the sea-bottom, and while part of the resulting débris is carried out to deeper water, another part is thrown upon the landward edge of the submarine cut, to form a submarine bar roughly parallel with the shoreline.

Where emergence is gradual, as is perhaps usually the case, the bar may form before the mainland is appreciably cliffed by wave action, and the nip observed later may then be wholly the work of lagoon waves. When the bar has been built upward to the water surface, the shoreline of emergence may be said to have passed its initial stage and to have entered that of youth.

Young Stage. — As soon as the submarine bar lying offshore has been raised above the surface of the water, we can distinguish an outer and an inner shoreline; the first bordering the seaward side of the *offshore bar*, or barrier beach as it is often called, while the second is the original shoreline, now characterized by the low cliff or nip bordering the edge of the mainland. Between the mainland and the bar lies a *lagoon*, on whose surface small waves only can be generated, both because of the shallow depth and the comparatively short stretch of open water exposed to wind action. Inasmuch as the offshore bar is the most striking feature of the young shoreline of emergence, we may appropriately consider the precise method of its development somewhat fully.

Offshore Bar. — Various theories have been offered to account for the production of a narrow bar lying parallel to, but some distance from, a gently sloping sandy shore. One writer has even gone so far as to deny their marine origin. Bryson[1], writing in 1888, considered that the offshore bars along the south side of Long Island had been produced by subglacial streams, and naïvely remarks: " These beaches have generally been held to be of marine origin, but this idea is being abandoned." In a later paper[2] he states that the bars are really kames. We can at least agree with his admission that " this hardly seems possible." Schott[3] tried to explain the remarkable offshore bar bordering the north shore of Yucatan as the product of outward pressing land waters meeting the resistance of the sea.

Louis Agassiz[4] suggested that at least along the coast of the southern United States the offshore bars of sand rested upon pre-existing coral reefs. Merrill[5] was convinced that these bars were " formed under water by wave and current action," but experienced difficulty in accounting for the appearance of their crests above the surface of the water. He solved the problem by assuming an elevation of the sea-bottom which

"brought these sand-bars above water into a horizon of æolian action. Once above the sea, the beaches would maintain their existence." McGee[6], on the other hand, seems to have regarded the presence of offshore bars and keys as a proof of coastal subsidence, the sea having encroached upon the land so rapidly as a consequence of the sinking movement that the bars were left behind. This implies the belief that such bars begin to form at the edge of the mainland, which is clearly the conception of Ganong[7] who writes as follows concerning small bars off the coast of New Brunswick: "They no doubt formed against the margin of the flat upland as ordinary shore beaches. But the steadily progressing subsidence carried the land beneath the sea faster than the beaches, whose rate of inward movement is largely determined by the rate of erosion of the protecting headlands, could follow; hence the lagoons were formed." While the forms described by Ganong should perhaps be classed as bay bars, the principle involved does not differ from that in the case of the offshore bars called "keys" by McGee. It would seem that a similar idea as to the origin of offshore bars has been entertained by David White and C. A. Davis, as it is otherwise difficult to understand their belief that such bars should be regarded as proofs of coastal subsidence[8].

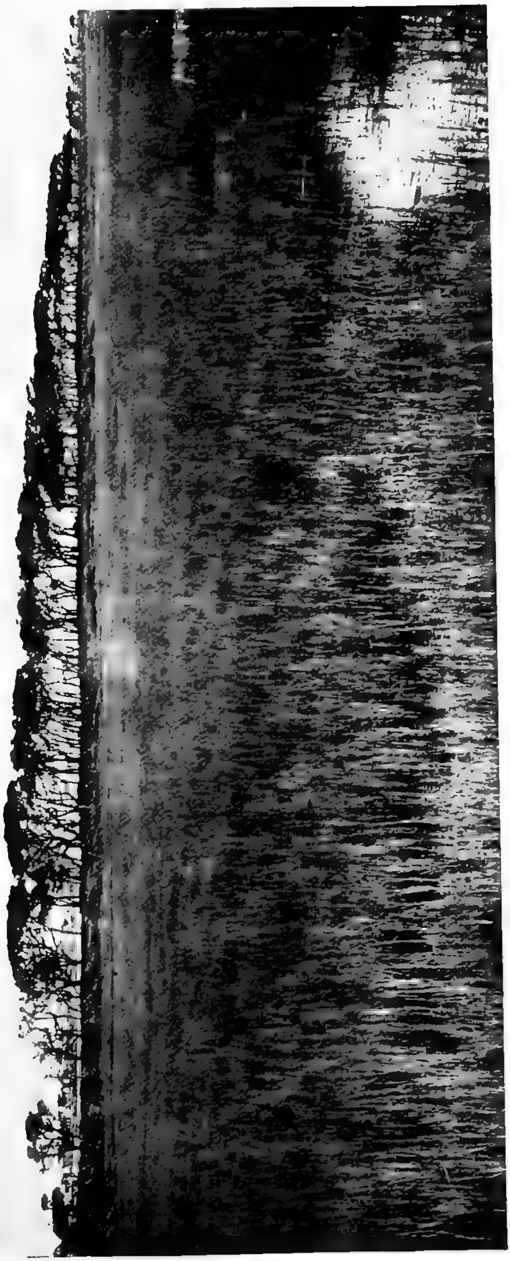
One of the best accounts of offshore bars is of earlier date than any of the discussions mentioned above, having been published by Élie de Beaumont[9] in 1845. In his "Leçons de Géologie Pratique" this keen observer not only describes the bars at much length and explains how wave action on a shallow bottom removes part of the material and heaps it up in a ridge parallel to the shore, but also states that this change involves a readjustment of the submarine slope to bring it into closer harmony with the movements of the water. In other words, he recognizes the effort of the sea to establish a profile of equilibrium, and that the offshore bar is one result of this effort.

Shaler[10] emphasizes the relation of offshore bars to shorelines of emergence, and assumes an uplift of the continental shelf as the first step leading to the formation of such a bar. The second step is considered by him to be shallowing of the offshore zone by deposition of débris eroded from the margin of the land, and other débris moved landward by the friction of the waves upon the bottom farther seaward. Not until this shal-

lowing has occurred does he imagine bar formation to begin. Storm waves then break at a considerable distance from the land, and drop the débris they were moving landward, thus building a ridge parallel with the shore which is permanently preserved in case it rises above the surface of the sea.

Gilbert[11] and Russell[12] do not appear to make a clear distinction between bay bars and offshore bars. Thus Gilbert's description of what he terms " the barrier " would seem to apply to offshore bars formed in gradually shallowing water. This interpretation is sustained by the fact that he compares with " the barrier," those " low ridges of sand or gravel running parallel to the shore and entirely submerged " which can be traced continuously for hundreds of miles along the shores of Lake Michigan, but whose origin is uncertain[13]. On the other hand, the bay bar at Stockton, Utah, is called both a " bay bar " and a " barrier[14]; " and the dependence upon shore drift ascribed to " barriers " would seem more characteristic of bay bars than of offshore bars. Russell[15] describes the formation of " barrier-bars " in terms which recall Gilbert's description of " the barrier "; and compares them with the submerged ridges paralleling the Lake Michigan shores. But Russell's illustration of " barrier-bars " shows ordinary bay bars closing the mouth of a small bay.

Assuming that Gilbert and Russell intended their descriptions to apply equally to offshore bars and bay bars, and taking Gilbert's description for examination as being the more complete of the two, we may next note the essential elements of this theory of offshore bar formation. According to Gilbert the material of which the bar is composed consists of "shore drift " which is being moved parallel to the coast by longshore currents. On a gradually shallowing shore " the waves break at a considerable distance from the water margin. The most violent agitation of the water is along the line of breakers; and the shore drift, depending upon agitation for its transportation, follows the line of the breakers instead of the water margin. It is thus built into a continuous outlying ridge at some distance from the water's edge. . . . The barrier is the functional equivalent of the beach. . . . The beach and the barrier are absolutely dependent on shore drift for their existence. If the essential continuous supply of moving detritus is cut off, . . . the structure (is) demolished by the waves which formed it[16]."



Former offshore bar near Wrightsville, North Carolina, rising above the marsh surface back of the present offshore bar.

Davis follows Shaler in relating the offshore bar to a shoreline of emergence, but does not admit the necessity of shallowing by deposition before bar formation can commence. He follows de Beaumont and Shaler in deriving the material of the bar from the offshore bottom, and disagrees with Gilbert who regards the material of the bar as shore débris in process of transportation parallel to the shore; for while Gilbert believed longshore transportation to be absolutely necessary, Davis states his conviction that offshore bars "might be developed essentially under the control of on- and offshore action alone[17]." The successive stages in the development of an offshore bar are described by Davis at some length in his "Erklärende Beschreibung der Landformen," where the discontinuous character of the bar during its initial stage, and the progressive narrowing of tidal inlets to a limiting size determined by an ultimate equilibrium between tidal and longshore currents, are emphasized[18]. Shaler[19], on the other hand, believed that the offshore bar had great continuity when first formed and that the so-called tidal inlets were really "outlets" formed by the bursting through of land waters dammed off from the sea by the bar.

Agassiz's theory, connecting offshore sand bars with coral reefs, may be dismissed on the ground that records of numerous wells drilled on the offshore bars along the coast of the southeastern United States fail to show the presence of such a reef below the sandy surface. While it is true that coral limestone sometimes underlies a ridge of beach or dune sand, as for example in the Florida keys, such a relation is not typical for the offshore bars from Long Island and New Jersey to Texas. Both theoretical considerations, and direct observations of small offshore bars raised above the level of lakes by wave action alone, justify us in rejecting Merrill's contention that an elevation of the sea-bottom is necessary to bring the bar crest above water. Equally untenable is the position of McGee, Ganong, White, and C. A. Davis that a subsidence of the sea-bottom is necessary for the development or maintenance of offshore bars. As this conclusion is of much importance in connection with the problem of recent coastal subsidence, we will return to it in a later paragraph. That portion of Shaler's statement which calls for offshore deposition of wave-eroded débris before bar formation can begin,

seems unnecessary; for simple uplift of a very gently sloping sea-bottom will produce the shallow offshore bottom which alone is necessary for the application of the theory of bar formation which he supports. The opinion that tidal inlets are really outlets formed by land waters bursting through a formerly more or less continuous bar, an opinion expressed by others[20] besides Shaler, is not supported by the evidence. Inlets are continually being opened through offshore bars and through bay bars which are already so discontinuous as to make the damming of land water an impossibility. The forcing of the opening from the seaward side by wave attack has repeatedly been observed; and the sudden rise of water in the lagoon immediately after the breaching of the bar, as at Scituate during the storm of 1898, proves that the sea, and not the land waters in the lagoon, may be the higher. Occasional inlets may be opened from the landward side; but as a rule the beach is forced by the waves of the sea.

The theories of de Beaumont and Gilbert seem most worthy of critical consideration. It does not seem necessary to rely upon ordinary " shore drift " either for the initiation or maintenance of offshore bars, as is required by Gilbert's theory. There is, to be sure, abundant evidence of longshore transportation of débris on the seaward side of most offshore bars; but it seems impossible to assign the vast volumes of material in the great bars along the south Atlantic and Gulf coasts of the United States to a source at one or the other end of such bars where they may connect with the mainland, or may recently have done so. The supply of débris from headlands is so small, and the loss of material from attrition under wave action along the face of the bars must in the aggregate be so large, that notwithstanding the impossibility of making a reliable comparison between these two factors, one is impressed with the probability that the bars would suffer rapid destruction were some other source of supply not available. An adequate source, both for the initial building of the bars and for their maintenance during a slow landward migration, is furnished by the shallow sea-bottom; and the on- and offshore action of waves is alone sufficient to excavate this material and build it into bars. That some material is also furnished by longshore transportation from the bases of cliffed headlands, and that material eroded from the sea-bottom suffers longshore transportation, is not to

be doubted. Such action must, however, be regarded as incidental and not vital to offshore bar formation.

Deductive Study of Offshore Bar Profiles. — The fact that Gilbert's theory of offshore bar formation does not imply erosion of the sloping sea floor, whereas de Beaumont's theory requires such erosion, suggests that the difference in profiles expectable in the two cases might enable one to determine which of these two most promising theories is best able to explain existing offshore bars. In other words, it occurred to the writer that the actual profiles of present-day offshore bars should clearly indicate the effects of extensive bottom erosion if de Beaumont's theory be correct, whereas such pronounced evidence of bottom erosion should be lacking if the bars formed according to Gilbert's theory. I therefore suggested this problem to Miss Bertha M. Merrill, a graduate student in physiography at Columbia University, as one which might yield tangible results. In the following paragraphs I have, with her permission, drawn freely upon her report of profile studies.

It may be noted that de Beaumont's theory does not exclude the possibility of some longshore transportation of débris by current action, although it necessarily implies that such action must be of minor importance. Débris cut from the original sea-bottom is sufficient to form the bar, and is assumed to be the principal source of supply. Gilbert's theory would seem on first reading to exclude all erosive action of onshore waves; but it is doubtful whether that author would altogether deny a minor rôle to débris eroded from the sea-bottom by the waves, and by them contributed to the growing bar. The essence of Gilbert's theory is that the bar absolutely depends for its existence upon, and is therefore largely composed of, débris brought from a distance by longshore currents. It would appear, therefore, that the profiles established by either of the two methods of bar formation operating alone might be slightly modified by the minor co-operation of the other method; but that such modifications would be so slight as not materially to change the essential nature of the profile characteristic of each method.

It will be convenient to consider first the profiles expectable on the basis of Gilbert's theory. Figure 90 shows the profile of a partially emerged coastal plain near the shore of which a

bar (b) has been built upon the uneroded sea-bottom through deposition by longshore currents. Because the bottom has not been eroded, the projection of the sea-bottom slope (ss') will intersect the sealevel surface at the inner edge of the lagoon (l). Even if the land area be dissected subsequent to uplift, the pro-

Fig. 90.

jection of the sea-bottom slope will still intersect the sealevel surface at the inner edge of the lagoon, although it will no longer coincide, as in the initial stage, with the land surface.

In case the sea-bottom is aggraded in the vicinity of the bar, but decreasingly so seaward from the bar, the projection of the

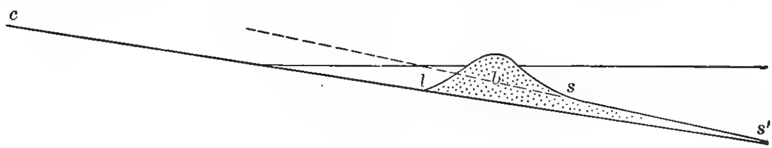


Fig. 91.

aggraded sea-bottom slope (ss', Fig. 91) will intersect the sealevel surface *seaward* from the inner edge of the lagoon.

We may imagine a third case in which the sea-bottom is aggraded in the vicinity of the bar, but to an increasing extent as one goes seaward. Then the projection of the aggraded sea-

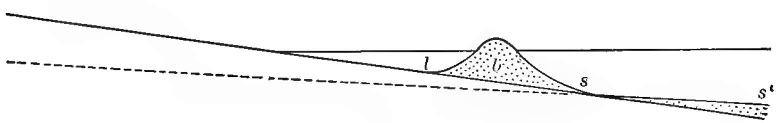


Fig. 92.

bottom slope (ss', Fig. 92) would intersect the sealevel surface *landward* from the inner edge of the lagoon. This case is highly improbable, for, according to Gilbert's theory, the bar is built up in the zone of maximum wave agitation. This zone occurs where the greatest number of waves expend their maximum energy upon the sea-bottom. Seaward from the bar, agitation is less because fewer waves are large enough to break there.

Since deposition is dependent upon, and proportional to, the amount of agitation, deposition decreases gradually away from the bar. Hence it is difficult to conceive an area of maximum deposition at *b*, an area of little or no deposition at *s*, and an area of increasing deposition at *s′*.

We conclude that in all profiles expectable according to the Gilbert theory, *the sea-bottom slope projected will intersect the sealevel surface at or seaward from the inner margin of the lagoon.*

Let us next consider the profiles which might characterize offshore bars constructed according to the de Beaumont theory. Figure 93 shows such a profile in which the original slope of a partially emerged coastal plain (*cc′*) has been eroded by the

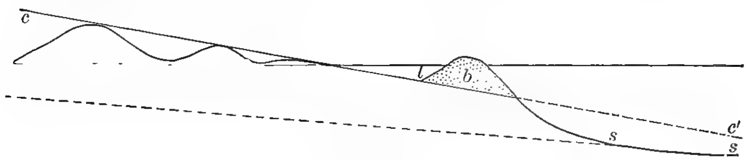


FIG. 93.

waves to produce a new sea-bottom (*ss′*), while a portion of the débris has been thrown up into an offshore bar (*b*). It appears that the projection of the sea-bottom slope (*ss′*) will intersect the sealevel surface some distance landward from the inner margin of the lagoon. Such a case would occur when all the material cut from the sea-bottom was either piled up in the bar

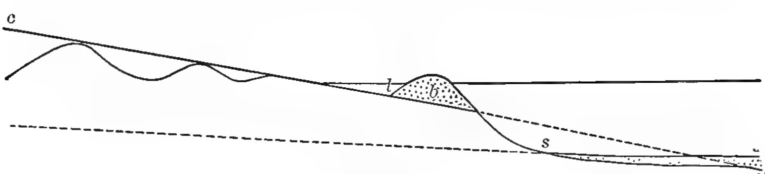


FIG. 94.

or carried too far seaward to affect this portion of the profile. A more probable profile is that represented in Figure 94, which shows the sea-bottom aggraded by deposition of part of the erosion products (*s′*) seaward from the zone of maximum wave attack (*s*). Again the projection of the sea-bottom slope (*ss′*) will intersect the sealevel surface some distance landward from the inner margin of the lagoon.

In both the above cases we have imagined that the angle of slope of the initial coastal plain and its seaward continuation is greater than the angle of slope of the newly fashioned sea-bottom. It is conceivable, however, that the original slope of the coastal plain might be so extremely gentle that the new submarine slope would be appreciably steeper. Such a condition is represented in Figure 95, from which it will be seen that in cases of this kind the projection of the sea-bottom slope (*ss'*)

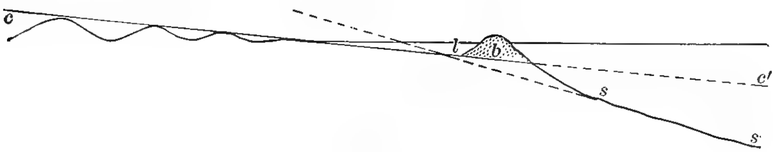


FIG. 95.

may intersect the sea-level surface *at* or *seaward* from the inner margin of the lagoon. The situation would be the same, of course, were the more steeply sloping sea-bottom a surface of aggradation, as shown in Figure 96. It would seldom happen that the projected sea-bottom slope would emerge exactly at

FIG. 96.

the inner margin of the lagoon. It should be noted that in cases of this kind the very gentle initial offshore slope will cause waves to break far from land, and the resulting bar will enclose a lagoon of exceptional width.

We conclude, therefore, that in profiles expectable according to the de Beaumont theory *the sea-bottom slope projected will intersect the sealevel surface landward from the inner margin of the lagoon*, except in those cases where the original surface slope is exceptionally low.

We may summarize the results of the deductive study of profiles as follows:

Class I. If the sea-bottom slope projected intersects the sealevel surface *at* the inner margin of the lagoon, the offshore bar was probably formed according to Gilbert's theory.

Class II. If the sea-bottom slope projected intersects the sealevel surface *landward* from the inner margin of the lagoon, the bar was probably formed according to de Beaumont's theory.

Class III. If the sea-bottom slope projected intersects the sealevel surface *seaward* from the inner margin of the lagoon, the bar may have been formed according to either theory; where the seaward slope of the land is at all pronounced, probabilities favor the Gilbert theory; where the coast is unusually flat and the lagoons very broad, the de Beaumont theory may apply.

Comparison of Actual Profiles of Offshore Bars. — To test the merits of the two theories, eighteen profiles were constructed for coasts having well-developed offshore bars. For this purpose the United States Coast and Geodetic Survey charts and the United States and Dutch Hydrographic charts were used. In order to eliminate from the profiles local and minor irregularities of the submarine slope, all the soundings within a zone of certain width, varying from five to seven miles according to circumstances, were projected on a single vertical plane normal to the shoreline, and the mean curve taken as the profile for that zone. Because such bars appear in great perfection off our own Atlantic and Gulf coasts, and because these coasts have been thoroughly charted, a majority of the profiles were taken from these regions. The others were constructed across bars of the North Holland, German, and Venetian coasts.

The results for each case, with appropriate comments, are briefly presented below:

Figure 97, Profile through the Gulf of Venice. From United States Hydrographic Chart, Adriatic Sheet I. The sea-bottom slope projected (broken line) intersects the sealevel surface landward from the inner margin of the lagoon, thus placing the profile in Class II.

Figure 98, Profile through the Kurische Nehrung and Haff on the Baltic coast. From United States Hydrographic Chart, Baltic Sheet II. The sea-bottom slope projected again intersects the sealevel surface landward from the inner margin of the Haff, showing that this profile also belongs in Class II.

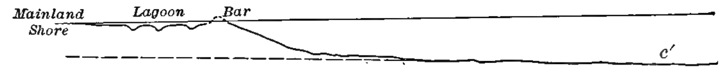


Fig. 97.

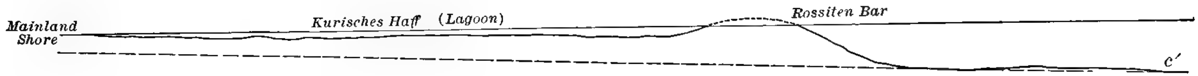


Fig: 98.

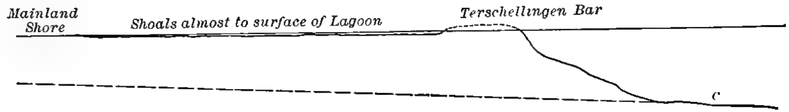


Fig. 99.

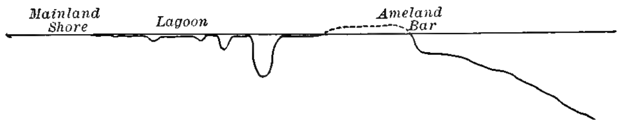


Fig. 100.

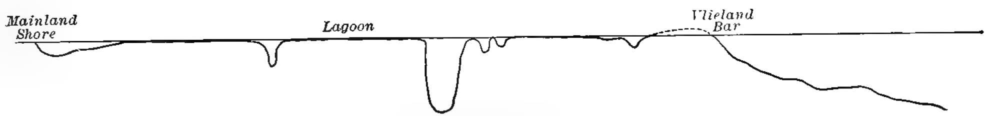


Fig. 101.

Profiles through offshore bars. Horizontal scale, 1 inch = about 6 miles; vertical scale, 1 inch = a little less than 400 feet. See text for descriptions.

Figures 99, 100, 101, Profiles through the Terschelling, Ameland, and Vlieland bars on the North Holland coast. From Dutch Hydrographic Charts Nos. 205 and 224. The profile through the Terschelling bar clearly belongs in Class II. In the case of the Ameland and Vlieland bars, too little of the sea-bottom slope is shown on the chart to serve as a basis for projection; but from the relation of these areas to the Terschelling area, and from other data for the sea-floor topography, it is known that both profiles belong in Class II.

Figures 102, 103, Profiles through the Cape Hatteras bar, North Carolina coast. From United States Coast Survey Charts Nos. 1232 and 1229. The coast is " an excessively flat plain," and the lagoon exceptionally wide. Both profiles belong in Class III.

Figure 104, Profile through Currituck Beach, northern coast of North Carolina. From United States Coast Survey Chart No. 1229^{II}. This part of the coast is less flat and the lagoons correspondingly narrower than further south where the profiles shown in Figures 102 and 103 were taken. The profile through Currituck Beach unquestionably belongs to Class II, as shown by Figure 104.

Figure 105, Profile through Assateague Island bar, Maryland. From United States Coast Survey Chart No. 1220. The profile appears to show local submarine bars, possibly of the low and ball type discussed later, and clearly belongs to Class II.

Figures 106, 107, Profiles through the offshore bar of the New Jersey coast, near Barnegat Inlet. From United States Coast Survey Charts Nos. 121 and 122. Both profiles belong in Class II.

Figure 108, Profile through Fire Island bar, south coast of Long Island, New York. From United States Coast Survey Chart No. 1214. The profile belongs in Class II.

Figure 109, Profile through Galveston Bay and Bolivar bar, Texas coast. From United States Coast Survey Chart No. 204. The profile belongs in Class II.

Figure 110, Profile through Matagorda Bay and bar, Texas coast. From United States Coast Survey Chart No. 207. The profile belongs in Class II.

Figures 111, 112, 113, and 114. Profiles through Laguna Madre and Padre Island bar, Texas coast, in latitudes 27° 25′, 26° 10′, 26° 45′ and 26° 25′ respectively. From United States Coast

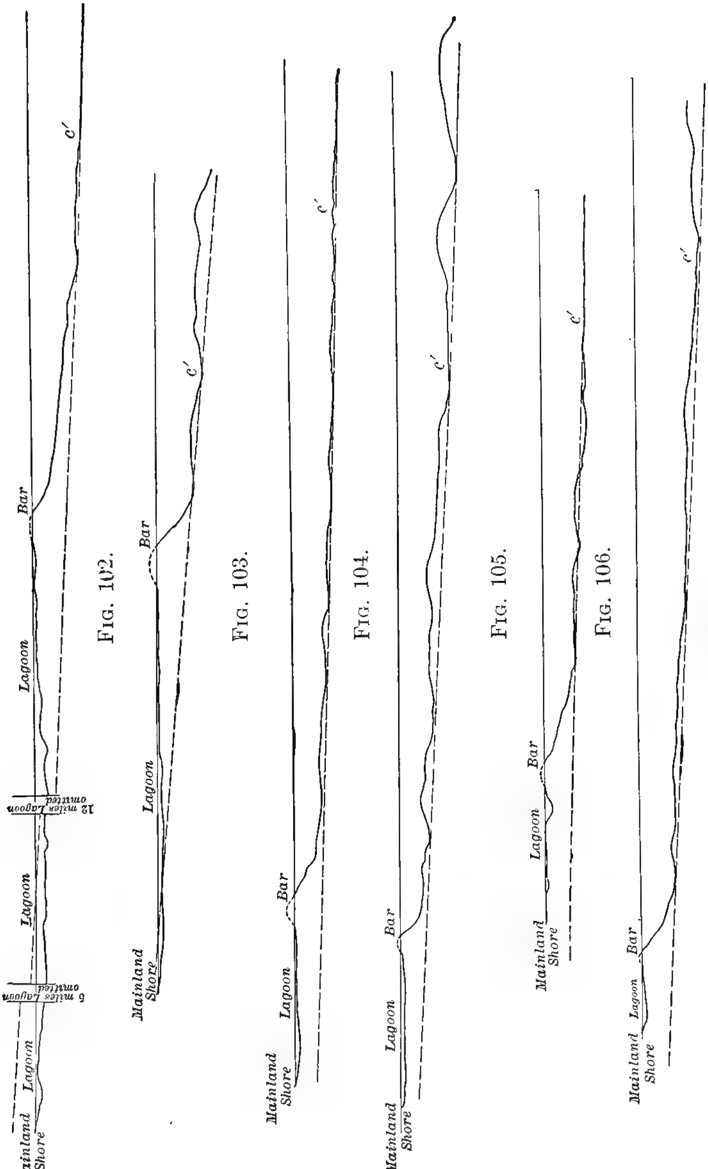


FIG. 102.

FIG. 103.

FIG. 104.

FIG. 105.

FIG. 106.

FIG. 107.

Profiles through offshore bars. Horizontal scale, 1 inch = about 6 miles; vertical scale, 1 inch = a little less than 400 feet. See text for descriptions.

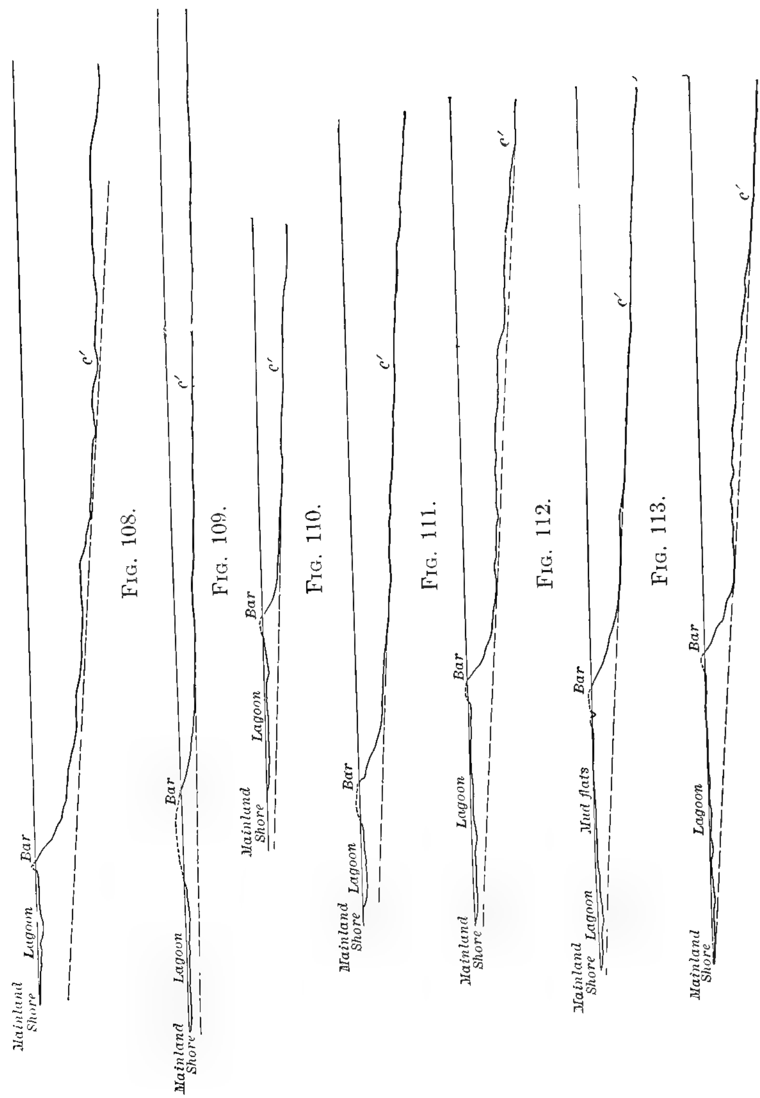


FIG. 108.

FIG. 109.

FIG. 110.

FIG. 111.

FIG. 112.

FIG. 113.

FIG. 114.

Profiles through offshore bars. Horizontal scale, 1 inch = about 6 miles; vertical scale, 1 inch = a little less than 400 feet. See text for descriptions.

Survey Charts Nos. 210, 211, and 212 respectively. The profiles are arranged according to increasing breadth of lagoon. The first three clearly fall in Class II; and the fourth, where the lagoon is exceptionally broad, appears to do so, although it closely approaches the conditions of Class I.

Summarizing the results obtained from the foregoing examination of profiles through offshore bars, we note that out of eighteen profiles studied, sixteen fall in Class II, although one of these approaches closely the conditions of Class I. The two remaining profiles fall in Class III. Both the two profiles of Class III and the profile closely approaching Class I occur off very flat coasts where the lagoons are exceptionally wide, as would be expected were the bars formed according to the theory of de Beaumont. In other words, fifteen of the profiles certainly fall in Class II, indicating that the bars were formed according to the de Beaumont theory; while the remaining three profiles, explicable according to either the Gilbert or the de Beaumont theory, show features suggesting that they also were formed according to the de Beaumont theory.

It might be argued that the bars first formed according to the Gilbert theory and were then pushed landward, the waves cutting into the sea-bottom and adding part of the erosion products to the bars. This would be to assume an initial stage of bar formation the validity of which could not be tested by appropriate facts of observation, and to admit that the bars as we now see them owe their existence, in part at least, to the process outlined by de Beaumont and more fully described by Davis. Under these circumstances it is perhaps more reasonable to accept the de Beaumont theory of bar formation, not forgetting, however, that longshore transportation of débris is an accessory process of very great importance.

Development of the Offshore Bar. — In tracing the development of an offshore bar we may therefore imagine a gradually shallowing sea-bottom on which small waves break at the initial shoreline and excavate a marine cliff and bench, while large waves break farther out and proceed to excavate the same forms in the offshore bottom. Along the outer zone part of the excavated material is deposited just landward of the breakers, in less agitated water; that is, on the crest of the submarine cliff. As the waves excavate deeper and farther landward the deposit on the

summit of the submarine cliff increases in volume until a submarine bar of significant height, and indefinite length parallel to the inner shoreline, is formed. Further growth brings the crest of the bar above water at irregular intervals, giving a chain of islands separated by wide spaces of shallow water covering the still submerged portions of the crest. With continued excavation along the seaward face of the bar and addition to its crest, the islands increase in number and in length, progressively narrowing the water spaces between them and ultimately coalescing to a greater or less extent to form a more nearly complete barrier between the open sea and the shallow lagoon

Tidal waters which formerly ebbed and flowed across the wholly submerged bar with little hindrance, now find themselves confined to a limited number of increasingly narrower passageways between the ever lengthening above-water portions of the bar. As the openings decrease in size, the tidal currents (including the all-important hydraulic currents generated by tidal action) flowing into and out of the lagoon increase in velocity. They compensate in some measure for the increasingly restricted breadth of their passageways by cutting deeper channels across the still submerged portions of the bar; and it seems probable that this process may often be carried so far that tidal channels are cut clear through the bar and into the original sea-bottom below.

As the submarine bar approaches the surface it comes more and more under the influence of the local wind-generated waves which affect the water to a shallow depth only. As a majority of these waves strike the seaward face of the bar obliquely, beach drifting alongshore becomes increasingly important, and soon is the dominant factor in the narrowing of tidal inlets. No longer are the above-water portions of the bar extended and the inlets narrowed mainly by simple vertical upbuilding of the still submerged parts of the bar. Instead, the débris eroded from the bottom and cast up against the face of the bar is attacked by oblique wind-made waves and transported laterally to be deposited at the ends of the elongated islands, thereby increasing their length and narrowing the inlets. This action is directly opposed to that of the tidal currents which pass in and out of the inlets and endeavor to keep them open by removing material deposited by the longshore currents. So long

as the longshore action is dominant, the inlets continue to narrow; but this very narrowing, by confining the tidal currents to smaller and smaller cross sections, progressively increases their velocity. A time must come when the inlets are narrowed enough to give the tidal currents a strength equivalent to that of the longshore currents. Thereafter deposition at the margins of the inlets by longshore currents is followed by equivalent erosion through the agency of tidal currents. Equilibrium between the two opposing forces is established, and the breadth of the inlets remains approximately constant.

The required breadth may be maintained by a few comparatively broad inlets, or a larger number of narrower inlets. Since a larger tidal range means stronger tidal currents, we should expect to find some relation between the range of the tide along a given coast and the number or size of the inlets through its offshore bars. Such a relation seems to exist. Thus along the New Jersey coast, where the tidal range is from 4 to 5 feet, inlets are more frequent than along the coast of Texas where with a tidal range of but 1 or 2 feet one offshore bar extends unbroken for about 100 miles.

Factors Controlling the Number and Breadth of Tidal Inlets. — It is commonly assumed that the amplitude of the tide is the only factor involved in determining the number and width of tidal inlets through offshore bars. Both theoretical considerations and field observations negative this assumption. In addition to the varying strength of longshore action (mainly beach drifting), the volume of land water, the extent to which the lagoon is filled with sediment or marsh deposits, the abundance and rapidity with which débris is supplied, and the strength of storm-wave attack, are all factors o importance. With the same tidal range along two offshore bars, it may happen that longshore current action is weak on one, but vigorous on the other. Under such conditions the one with the weaker longshore currents will have more or wider inlets. Where large rivers empty into a lagoon, the ebb current of the tide is greatly reinforced by the land waters, and will keep open inlets which would otherwise be narrowed or closed. As sedimentation and marsh growth decrease the water space of the lagoon, the volume of tidal waters admitted and the strength of the tidal currents is reduced, in consequence of which longshore currents may be

able to narrow or even close some of the inlets. If an abundance of débris is supplied to longshore currents with great rapidity, the closing of inlets will be more readily accomplished than if a smaller amount of débris is supplied very slowly. An inlet, once closed, might never be re-opened were it not for breaches made in the bar by storm-wave attack. Tidal action tends to keep inlets open; but, except in the case of an unusually high tide overflowing a low point on a bar, does not tend to produce inlets. Impounded land water may in rare instances open an inlet after the manner described by Shaler; but inlets are more commonly re-opened during exceptional storms by vigorous wave erosion. A bar exposed to the waves of an occasional great storm may thus be breached, where one less exposed would remain intact.

On the other hand, it matters little how many inlets may be opened by the waves, longshore currents will soon close all except those kept open by tidal currents reinforced by outflowing land waters. If the tidal range is such as to generate currents capable of maintaining two inlets of a given breadth through a certain bar, and storm waves cut two additional inlets, the tidal waters will for a time flow through the greater number of openings with decreased velocities. Longshore currents will therefore dominate the tidal currents at the inlets, until deposition has narrowed all of the inlets, or closed two of them (often the older ones), leaving the other two of the required breadth and thereby re-establishing a condition of equilibrium. Or, if a storm drives waves obliquely upon a coast in such manner as greatly to accelerate the longshore transportation of débris, all the inlets through a bar may be closed by excessive deposition in spite of tidal currents Once the inlets are closed, the tidal currents cease to exist; and the inlets will remain closed until storm waves or some other agency makes new breaches through the bar. In general we may say that waves tend to make inlets, tidal currents to preserve them, and longshore currents to close them.

Theory of Tidal Inlet Distribution. — That the supply of débris brought by longshore currents may be more important than differences of tidal range in determining the number of inlets through a bar, is apparent from a study of certain offshore bars which are supplied with débris derived from headlands to which the bar is at one end attached. Let us deduce the conditions

which theoretically should characterize offshore bar and lagoon development when the bar is attached to a headland, and longshore currents move from the headland toward the further extremity of the bar.

In the first place, it is evident that while wave currents may remove much material from the face of the bar and transport it seaward to deeper water, near the headland the loss may be more or less completely made good by new débris brought from the adjacent source of supply by longshore currents. The effect of this accession of débris is two-fold: the bar withstands the normal tendency of the waves to drive it landward since the waves have all they can do to take care of the new material continually being added to its face; and for the same reason the waves are less apt to cut inlets through the bar, while longshore currents utilize the abundant débris to seal up such inlets as may occasionally be formed. Accordingly we should expect a tendency for lagoons to be broad and bars to be continuous in the vicinity of headlands.

Toward that end of the bar most remote from the headland, conditions are very different. The débris from the headland has been ground fine in the course of its journey, and largely dissipated. Wave attack expends its full energy upon a bar which receives little material from the distant headland to offset the ravages of marine erosion. Hence the bar is driven landward with greater ease, and during its retreat the waves cut through first here, then there, forming inlets which are not closed as readily as where débris is more abundantly supplied. Far from headlands, therefore, there should be a tendency for lagoons to be narrow and for bars to be broken by frequent inlets.

We may deduce an interesting corollary as to conditions within the lagoon. Where the bar is continuous, little sediment from its seaward side can reach the lagoon, and that little must be brought in suspension by tidal waters entering by some distant inlet. Where inlets are abundant, more sediment can enter the lagoon with flood tide, even though this be the part of the bar most poorly supplied with débris from the distant headland. It must also appear that the end of the lagoon near the headland is least apt to have a constant salinity. At times the water may become nearly fresh, while high tides or tempo-

rary inlets will result in a high salt content. Such variations in salinity are unfavorable to the growth of either marine or fresh water vegetation. On the other hand, where numerous inlets keep the lagoon waters constantly salt, marine grasses thrive and contribute effectively to the filling of the lagoon. We conclude, therefore, that theoretically the "up-current" or headland end of a lagoon should be more open than the further end where marine sediment and marine vegetation unite to form a salt marsh filling.

The Theory Tested. — If we turn now to an examination of offshore bars and lagoons along the Atlantic Coast, we find that despite the manifest possibility of other factors complicating the situation, there exist substantial confirmations of the theory of inlet formation outlined above. (In the discussion which follows I have drawn freely upon the results of map studies made by Miss B. M. Merrill, under my direction.) On the south side of Long Island the longshore current moves westward along an offshore bar (Fig. 115) which is attached at its eastern end to a complex headland consisting of a terminal moraine and outwash plain. From Southampton, where the bar really springs from the mainland (it barely touches it at Quogue) westward to the Gilgo Lifesaving Station, a distance of 54 miles, there is only one inlet; in the next 22 miles, to Far Rockaway, there are three inlets. For sake of easy comparison with the cases which follow, we may say that nearest the headland the inlets occur at the rate of 2 to 100 miles, while farther away the rate is 14 to 100 miles. Great South Bay, the main lagoon, is wide and comparatively free from tide marsh in the half nearest the headland, narrower and almost filled with marsh in the farther half where inlets are frequent. The actual conditions are precisely those which deduction led us to expect.

The New Jersey coast is fringed by an offshore bar (Fig. 116) attached at its northern end to a headland consisting of the cliffed coastal plain between Long Branch and Bayhead. The longshore current moves southward from the headland. In the first 50 miles there are 2 inlets, in the next 50 miles, 8 inlets. In other words, nearest the headland the inlets average 4 to 100 miles, farther away 16 to 100 miles. As in the Long Island case the half of the lagoon nearest the headland has the greater average width and the smallest amount of marsh filling. Toward the

FIG. 115. — Offshore bar and lagoon of the Long Island coast, showing distribution of tidal inlets.

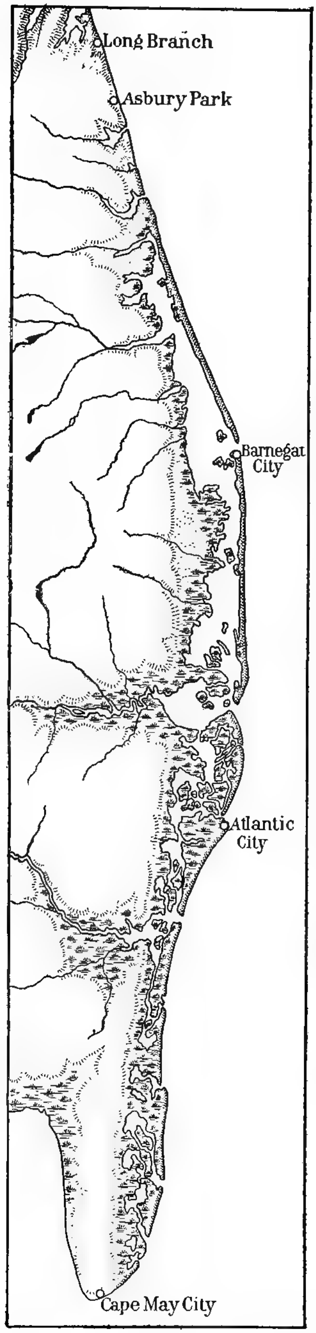


FIG. 116. — Offshore bar and lagoon of the New Jersey coast.

south, where inlets are frequent, we find the lagoon narrow and almost completely filled with marsh. In this case also the observed facts conform to the expectations as deduced from the proposed theory of inlet formation.

Similar conditions obtain off the coast between Delaware and Chesapeake Bays. An irregular headland extending from Cape Henlopen to Bethany Beach has attached to it and extending southward an offshore bar which continues for something over 50 miles before the first inlet is reached, whereas in the next 50 miles ten inlets occur. The relation is therefore roughly expressed by assigning a rate of 2 inlets per 100 miles for the headland end of the bar, and 20 inlets per 100 miles for the farther end. Near the headland end we have the open Chincoteague Bay. Farther south the lagoon is first narrow and marsh-filled; but the expectable relations are then masked by a widening of the lagoon area possibly as the result of an exceptionally flat initial sea floor which permitted the bar to form far offshore. It should be noted, however, that even here the inlets are close-spaced and the lagoon area largely filled by marshes or mud-flats, as required by the theory.

The Carolina coast is so complicated by the three cuspate bars forming Capes Hatteras, Lookout, and Fear that one might scarcely expect to find the relationships characteristic of simple offshore bars. Yet if we compare different sections of the coast in a broad way, ignoring local abnormalities, we seem to see the working of the same laws controlling cases previously discussed. The headland for this section is the margin of the coastal plain of Virginia, south of Cape Henry, and the shore currents move in a general north to south direction. We may recognize four natural subdivisions of the coast: a first section from the headland to Cape Hatteras, a second between Capes Hatteras and Lookout, a third between Capes Lookout and Fear, and a fourth between Cape Fear and a point just west of Little River, beyond which the offshore bar seems to touch the mainland again. In the first section the inlets number but 2 in a distance of 113 miles, and the lagoon attains a great width with comparatively little filling. The abnormal width in parts of the first two sections is probably due to an exceptionally gentle slope of the seafloor along the Cape Hatteras axis. In the second section of 72 miles, there are three inlets, giving an average

spacing of 4 to 100 miles, and the lagoon becomes comparatively narrow toward Cape Lookout. In the third section the number of inlets increases to 9 in 100 miles, while the lagoons narrow still more and become much more filled with marsh deposits. At Cape Fear the lagoon broadens out considerably, but the width here is only seven and one-half miles as compared to twelve and one-half at Cape Lookout, or thirty miles at Cape Hatteras. In the fourth section there are eight inlets in 40 miles, which is equivalent to a spacing of 20 inlets to 100 miles; the bar is driven back nearly to the mainland, and the narrow lagoon is almost completely filled with marsh. Despite its complexities the Carolina case appears to meet the requirements of the theory.

The Florida offshore bar is so complicated by the presence of hard coquina along some of its parts, by the complex cuspate foreland of Cape Canaveral, and by coral reefs farther south, that it does not properly come within the scope of our enquiry. If we consider the Texas coast, however, taking the Rio Grande delta as the headland supplying the débris, and ignoring the Rio Grande and Brazos Santiago openings in the immediate vicinity of the delta, we find an offshore bar extending northward, in the direction of what appears from sand migration to be the dominant longshore current, more than 100 miles before the first inlet is encountered. Near the headland we have the Laguna Madre, broadly open, but shallow because of the very gentle slope of the initial sea floor. Farther north the lagoon proper (no account should be taken of the drowned valley bays) grows narrower and the proportion of marsh filling increases.

In all of the cases described above there is a marked tendency for the number of inlets and the proportion of lagoon filling to increase, and for the width of lagoon to decrease, with increase of distance from headlands. This seems to confirm the theory that the amount of débris brought from headlands by longshore currents exercises an important control over the number of inlets through offshore bars, as well as upon the rate of bar retreat and lagoon filling. It should be noted, however, that in each case the tidal range increases from the headland toward the farther end of the bar, although the amount of increase between sections of no inlets or few inlets, and sections of numerous inlets is sometimes so slight as to be of doubtful importance.

Both distance from headland and range of tide co-operated to produce the observed results, but it is believed the former factor is the more important of the two.

Tidal Deltas. — The hydraulic currents generated by the tides, and called tidal currents in the foregoing paragraphs in conformity with well-nigh universal custom, produce certain features at the inlets which deserve brief notice. Débris brought by beach drifting or other longshore currents is seized by the inflowing or outflowing current at the inlet and transported into the lagoon or out to sea. Most of the débris is not carried far before being deposited in the quieter water of the larger water-body to form a *tidal delta*.[21] The typical tidal delta is wholly submerged and is double, one part facing landward and representing the result of deposition in the lagoon by incoming currents; the other part facing seaward and owing its construction to deposition in the sea by outflowing currents. Because the seaward part of the delta is exposed to the action of waves and longshore currents it is commonly stunted in its growth and margined by contours of simple curvature; only that portion in the lagoon is apt to acquire appreciable size and the lobate form of ordinary deltas (Fig. 117).

Migrating Inlets. — The exposure of most beaches to wave attack is such that longshore current action (usually beach drifting) in one direction predominates over that in the other. This results in a marked tendency for inlets to migrate in a certain definite direction — that of the dominant current. Deposition on that side of the opening to which longshore currents bring abundant débris tends to narrow the inlet, whereas erosion alone is operative on the other side. An excess of deposition on one side, accompanied by erosion alone on the other, must result in a lateral migration of the inlet along the bar in the direction of the dominant current, while the breadth of the inlet remains unimpaired. On the New Jersey coast south of Barnegat Inlet the inlets through the offshore bar migrate southward, while north of this point the direction of inlet migration is northward.

The presence of a dominant current along an offshore bar broken by inlets results in the development of offsets and overlaps similar to those already described in connection with bay bars (Chapter VI). As these two features are fully explained in the connection cited, it will not be necessary to consider them at

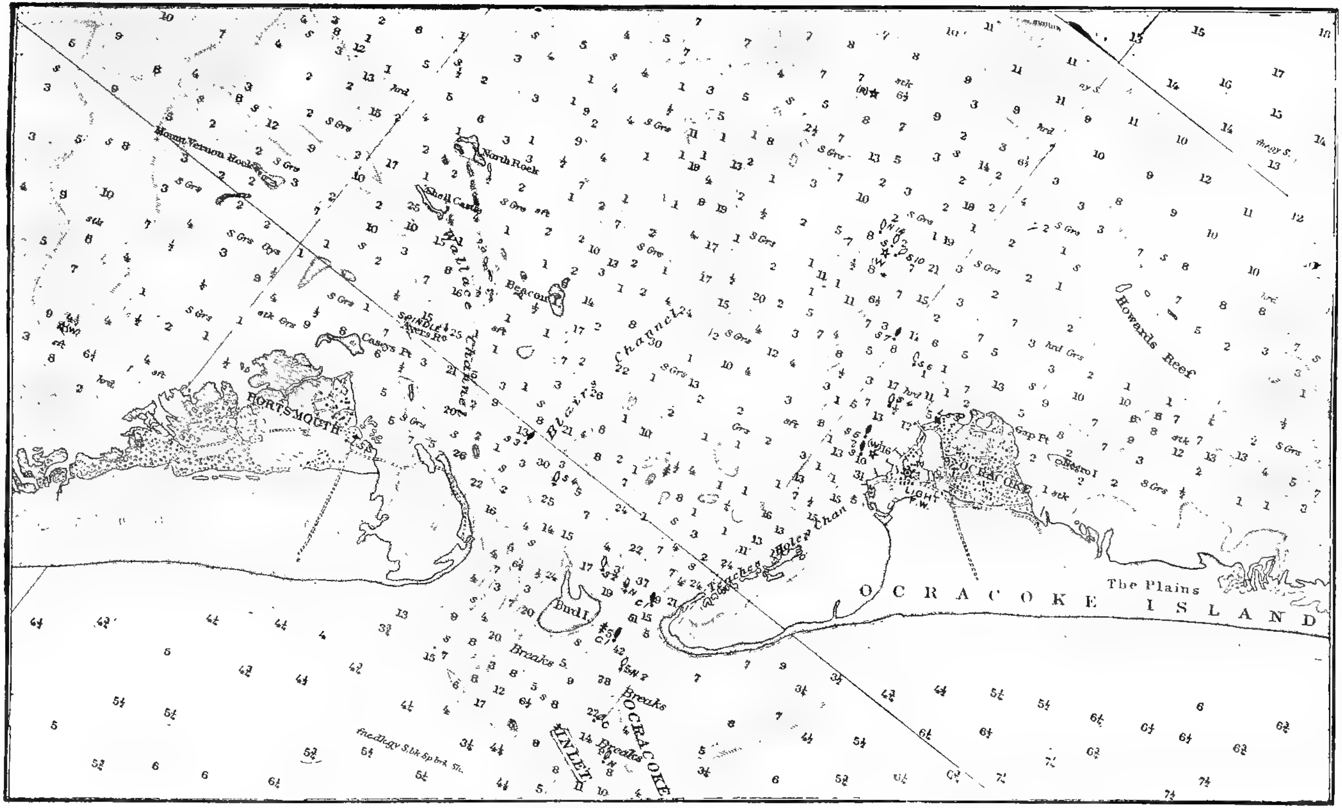


FIG. 117. — Tidal delta at Ocracoke Inlet, North Carolina coast. U. S. Coast Survey Chart.

greater length here. We may simply recall in passing that the southern New Jersey coast appears to afford an exception to Gulliver's rule[22] according to which the dominant current should "flow from the outer curve toward the inner one" along a shoreline marked by offsets. Here the direction of inlet migration proves that the dominant current is from the northeast; but according to Gulliver's rule the offsets at the inlets north of Cape May would require a current from the southwest. It is clear that the direction of offset may be determined in certain cases by some force other than the dominant longshore current.

One important consequence of inlet migration which seems not to have been fully recognized, will claim our attention when

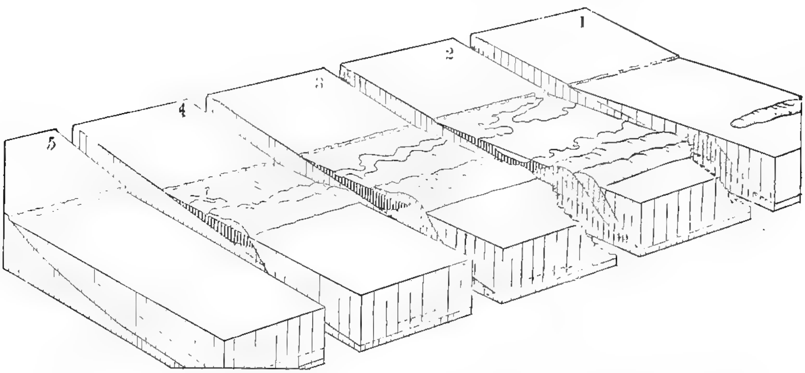


FIG. 118. — Stages in the development and retrogression of an offshore bar. (After Davis.)

we come to consider the landward retrogression of the offshore bar. It is quite generally assumed in descriptions of this last process that the bar comes to repose wholly upon the deposits of the lagoon or marsh as soon as it has moved landward a distance equivalent to its own breadth. This view is well exemplified by Figure 118, reproduced from a diagram given by Davis in his "Erklärende Beschreibung der Landformen[23]". It is quite evident, however, that the condition represented by this diagram could only obtain where inlet migration is either wholly absent or takes place slowly at the same time that bar retrogression is comparatively rapid. For the migration of an inlet along the bar results in the complete removal of that portion

of the bar and its underlying deposits toward which the opening is moving, down to the greatest depth reached by the tidal channel; while deposition on the up-currant side of the inlet forms an essentially new bar whose base rests, not on the surface of the lagoon deposits, but upon the erosion plane formed by the lateral migration of the inlet. Since the inlets probably reach to or below the original shallow sea-bottom in a majority of cases, the bar to leeward of the migrating inlet will commonly rest on the original sea-bottom deposits.

Inside the inlet the remains of the tidal delta left on the up-current or leeward side of the opening will prolong the landward side of the bar into the lagoon with a gentle slope; for just as successive deposits of sand on the up-current side of the inlet remain to form the new part of the bar, so the side of the delta away from which the inlet is migrating is progressively left behind to form a sheet of sand extending from the new bar out into the lagoon as a thinning wedge. If the offshore bar has a marsh behind it instead of an open lagoon, the result is essentially the same. Erosion will remove both the bar and the adjacent marsh deposits on the down-current or far side of the inlet, while deposition of sand on the up-current or near side will leave a bar resting on the eroded sea-bottom. This bar will be extended marshward by sand deposited along the near side of the tidal creek connecting with the inlet. A cross section through an offshore bar and marsh, after the bar had migrated toward the land a great distance, would in this case not look like stage 4 in Figure 118, as is generally assumed, but more like stage *G* in Figure 119.

In case the bar moves landward an appreciable distance after one inlet has migrated past the line of the cross section and before the next migrating inlet has reached that line, we would have the conditions represented in stage *F*, Figure 119, where the marsh deposits are exposed on the seaward side of the bar.

It is evident from the considerations just outlined that it may be difficult or impossible to determine how far landward from its original position an offshore bar has migrated. Were the assumed conditions of stage 4, Figure 118, commonly present after a considerable landward migration, the problem would be more simple; for soundings made through the marsh deposits would show an increasing depth of these deposits until the margin of

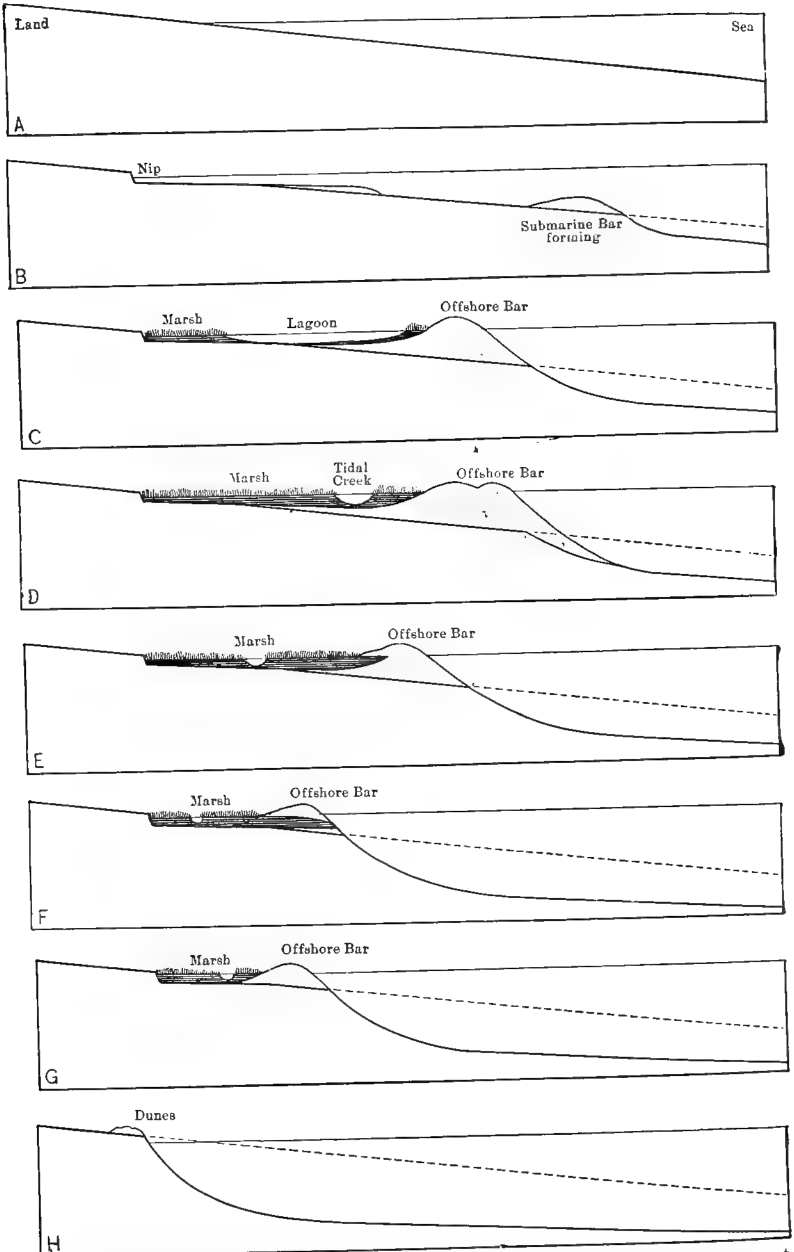


FIG. 119. — Stages in the normal history of an offshore bar, due account being taken of the effect of migrating inlets. Between stages *F* and *G* an inlet has migrated past the zone of the cross section, producing conditions similar to those in stage *C* or *D*.

the superposed bar was reached; and wells drilled on the bar would pass through a thick layer of marsh mud under the beach sands. On the other hand, soundings showing an increasing thickness of marsh deposits for some distance seaward from the inner shoreline, followed by a gradually decreasing thickness as the bar was approached (stage 2, Fig. 118), and records of wells on the bar showing that nothing but sand was encountered by drilling, would indicate that the bar was still in its original position.

Unfortunately such reasoning, although frequently followed, at least tacitly, when offshore bars are discussed, is not valid if shifting inlets are involved. It is clear from Figure 119, stage *G*, that the results of soundings and the well records accepted above as proving no landward migration of the bar, would be obtainable in the case of a bar which had really migrated far from its initial position. So also, soundings or well records might indicate only a slight landward progress of the bar, whereas the actual movement had been very great. Along the New Jersey coast, lines of soundings across the marshes show that beyond the axis of the marsh the peat and swamp muds thin out and the sandy bottom rises gradually toward the offshore bar. Well records frequently show that no marsh deposits were encountered in drilling, or that only small thicknesses of such deposits were found. Since the bar is repeatedly broken through by shifting inlets these facts cannot be regarded as evidence that the bar has changed but little from its former position, any more than the outcropping of small quantities of peat along the outer shorelines can be accepted as proof of an extensive landward migration of the bar. Either permanence of or marked change in the position of the bar must be proven, if at all, by other lines of evidence.

Lagoon and Marsh. — When the offshore bar is formed there is enclosed a long, narrow lagoon between the bar and the inner shoreline, the lagoon communicating with the open sea by means of the tidal inlets. Comparatively quiet water in the lagoon favors deposition of the fine débris which is derived from three principal sources. The products of attrition resulting from wave action on the outer surface of the bar are moved to the tidal inlets by longshore currents and the finer part is carried into the lagoon by tidal currents, to be widely distributed over

the shallow bottom; all the coarse material is added to the bar or dropped near the inlet to form the tidal delta. Rivers may bring sediment from the land surface into the lagoon, depositing the coarser part in the form of deltas along the inner shoreline, and delivering the finer part to the feeble currents in the lagoon for wider distribution. Onshore winds blow sand from the beach and dunes of the bar back into the lagoon. As a rule the coarser sand quickly drops into the water close to the lagoon shore of the bar, and only the finest material is wafted far over the surface of the waters before dropping into them to find a resting place on the submarine floor.

As material from these three sources accumulates, the bottom of the lagoon is built upward toward the surface. If the supply of fine sediment is unusually abundant the lagoon may eventually become filled with a deposit of almost pure clay or sandy clay, on the surface of which grow salt marsh grasses. Probably a more normal history would be something like that described by Shaler[24] in which eel-grass or other salt-water plants first gain foothold on the muddy bottom below low-tide level and aid the process of deposition by checking the currents passing through them. Later, as the lagoon bottom reaches a higher level, marsh plants are able to colonize the surface, and their remains may form no inconsiderable proportion of the completed deposit. The entire lagoon is thus ultimately filled with a clayey formation which includes, particularly in its upper portions, large quantities of vegetable matter; while its surface is covered with the grasses of a typical growing salt marsh.

Retrogression of Offshore Bars.—Just as continued wave attack ultimately forces the recession of other shoreline features, so the offshore bar must be driven landward in course of time. As previously explained the outer shoreline of the bar may temporarily be prograded; local disturbances of the shore profile of equilibrium may cause the bar to widen locally, as appears to be the case at Atlantic City; or general and long-continued excessive supply of shore débris may result in broadening a bar into a beach plain of great extent, such as that forming Cape Canaveral on the Florida coast. Occasionally after a bar is built the zone of bar construction is shifted so rapidly seaward that a broad swale or lagoon is left between the bar earlier formed and its later counterpart. If the swale or lagoon be occupied by

marsh, the first bar appears as a long ridge of dry land in the midst of the expanse of salt grass and water (Plate XLIV). But all such activities are temporary, and the time will come when loss of fine material from attrition and removal to deep water will exceed the diminishing supply of shore débris. The waves, relieved of the burden of excessive débris transportation, will then utilize their surplus energy in eroding the sea-bottom and driving the bar landward.

Material eroded from the face of the bar, and from the sea-bottom below, is hurled by waves to the bar crest or even beyond, and descends the black slope toward the lagoon with the assistance of over-wash from exceptionally high waves, and running water due to rainfall. This insures for the actively retreating bar a narrow breadth and an asymmetrical cross-profile, the front slope toward the sea being characteristically steeper than that toward the lagoon; while the lagoon shore is apt to show a series of rude deltas where overwash has projected beach material into the lagoon waters. If the lagoon has been replaced by salt marsh, the features are essentially the same, except that the overwash deltas spread out upon the marsh surface (Plate XLI), while the marsh muds and peat may become exposed below high tide on the seaward side of the bar.

Migrating tidal inlets tend to destroy all of the features just mentioned: the asymmetry of the bar profile, the overwash deltas, and the subjacent relation of marsh deposits to the bar. If the bar retreats rapidly while inlets are few and migrate slowly, the features described may be observed, except along that portion of the bar most recently re-formed. If the bar retreats slowly and intermittently, while inlets are numerous and migrate rapidly, the lack of symmetry and the overwash deltas may be poorly developed, while marsh deposits beneath the bar may be nearly or entirely lacking.

Gulliver[25] considers a prograding offshore bar as characteristic of the youthful stage of a shoreline of emergence, while a retrograding bar is the distinguishing feature of "adolescence." The exposure of marsh deposits on the seaward side of the bar is necessarily relied upon as the principal proof of retrogression. Davis[26] defines "late youth" as the period when the bar is driven landward far enough to show tide-marsh turf and mud on the outer side of the bar. One must doubt, however, whether

it is feasible to utilize such criteria as a basis for distinguishing different stages of shoreline development. In the first place it is usually impossible to tell from a map whether an offshore bar is advancing or retreating, so that maps would be of little or no use in determining whether a given shoreline was in youth or in its adolescent period (late youth). This difficulty is well exemplified in Gulliver's essay, where several shorelines of emergence are arbitrarily classified as "young," although the author admits that they may really be adolescent; but in three cases we read that "the scale of the map is too small to show indications in which direction the bar is moving," "whether advancing or retreating the writer does not know," and "the writer could find no evidence as to which way it is moving."

Even if field observations are available as an aid to classification, the case is little better. An offshore bar may be alternately retrograded and prograded due to changing conditions of equilibrium of the shore profile. It will hardly help us to assume that such a shoreline vibrates from youth to adolescence and back to youth again repeatedly. On the other hand, a bar which had been continuously but slowly retrograded for a long period of time might be erroneously assigned to the youthful stage in case migrating inlets removed the evidences of retrogression most commonly depended upon, such as the subjacent marsh deposits.

Normally an offshore bar should never prograde to any appreciable extent, but should retrograde from the moment of its initiation. Prograding implies a disturbance of normal conditions, a variation in the shore profile on one part of the coast due to abnormal activity of some one or more of the shore processes, as a result of which shore débris is supplied with exceptional rapidity to that part of the shore where prograding takes place. A possible exception to this statement is the presumably rare case in which progressively larger and larger storm waves built additions to the initial bar farther and farther seaward. It seems unwise to adopt as the criterion of "youth" a condition which has no sure place in the ideal normal history of shoreline development.

The three considerations set forth above force us to the conclusion that no great profit is to be derived from the attempt to distinguish different stages of shoreline development according

to whether the offshore bar is prograding or retrograding, whereas considerable confusion and misunderstanding is apt to result from such an attempt. We will therefore regard the offshore bar, with its associated lagoon or marsh, as characteristic of the youthful stage of the shoreline of emergence, making no attempt, in the present state of our knowledge of shorelines, to further subdivide the stage of youth. On this basis the New Jersey shoreline, from Bayhead to Cape May City, is a young shoreline of emergence.

Cuspate Offshore Bars. — Occasionally an offshore bar has a pronounced cuspate pattern. Such is the case at the Carolina Capes, on the offshore bar bordering North Carolina. Like ordinary cuspate bars, the cuspate form of offshore bars may be produced in a variety of ways. A favorite theory is that proposed by Abbe[27] for the Carolina Capes, and supported by Gulliver[28] and Davis[29] according to which the cusps result from deposition in the triangle of quieter water between two adjacent circling currents. A shoal or a former island some distance off a straight coast not infrequently produces a cuspate pattern in adjacent parts of an offshore bar. If the initial shoreline of emergence has pronounced projections or capes, then the offshore bar which is parallel to that shoreline will of necessity have a cuspate form imposed upon it. A study of the inner shoreline, back of the Carolina offshore bars, shows that the mainland itself possessed initial capes, later more or less cut back by wave action, which are perhaps fully competent to explain the Carolina cuspate bars.

Effect of Progressive Subsidence on Lagoon History. — If a shoreline of emergence bordered by an offshore bar is subjected to a gradual but continuous subsidence, certain departures from the normal history outlined above may be noted. Subsidence tends to deepen the water in front of the bar, thus enabling larger and more powerful waves to attack its face. This must result in an abnormally rapid retreat of the bar, since a bar moves landward just as fast as the water in front of it is deepened sufficiently to permit the near approach of large waves, whatever be the cause of deepening. If to the deepening performed by normal wave erosion we add a deepening due to progressive subsidence, certainly the landward movement of the bar will be appreciably accelerated. This does not mean

PLATE XLV.

Surface of a salt marsh near Boston, Massachusetts, which overlies a peat deposit 20 feet deep composed of remains of high-tide vegetation.

that the lagoon or marsh will be correspondingly narrowed, since subsidence causes the inner shoreline to encroach upon the land at the same time, and presumably at about the same rate that the bar moves landward. Both the bar and its associated lagoon or marsh advance upon the coast simultaneously. Migrating inlets, tidal deltas, and other shore phenomena develop as before. Sedimentation proceeds in the lagoon, but is not so apt to fill it as when the coast is stationary, since subsidence carries the bottom deposits downward and continually renews the water space which must be filled.

When a marsh has formed back of the bar, later subsidence, if not too rapid, may bring about several peculiar results. In the first place, as the surface of the marsh with its high-tide grasses is carried downward, new growths of grass continually arise upon the old in an effort to keep the marsh built up to the high-tide level (Plate XLV). The importance of this process was first recognized by Mudge[30] more than half a century ago, and has later been much emphasized by C. A. Davis[31]. The result is a deposit of salt marsh peat, composed of the roots and other remains of high-tide grasses, whose depth is an approximate measure of the minimum amount of subsidence. Sections through such a salt marsh, instead of showing high-tide grasses above, remains of eel-grass and other low-level grasses immediately below, and nearly pure silt or clay throughout the remaining depth of the lagoon deposit, as we should expect according to the Shaler theory of salt marsh formation, might show nothing but remains of high-tide vegetation from top to bottom, providing subsidence had progressed far enough to allow the offshore bar to move landward past the former position of the inner shoreline, and hence beyond the farthest limit of the initial lagoon deposits. As the salt marsh is progressively built upward it gradually encroaches upon the gently sloping surface of the subsiding mainland, overwhelming and burying the fresh-water vegetation which clothes that surface. Remains of the land vegetation may thus be preserved as a layer of fresh-water peat at the bottom of the salt marsh deposit, and may later be encountered in sections cut through the marsh to the solid ground below.

Another consequence of gradual subsidence after the marsh has formed is the complete disappearance of the nip along the margin of the mainland. So long as the lagoon persists, the

lagoon waters encroaching upon the subsiding mainland may be sufficiently agitated by winds to cut a small cliff at whatever level the water may stand. But after the marsh has once filled the lagoon area, there remains no force capable of cutting a straight cliff along the mainland shore, while the former wave-cut nip is carried downward under the marsh by subsidence and so lost to view. Thereafter the marsh surface and the gently sloping mainland surface intersect at a low angle which is often almost imperceptible.

Effect of Progressive Elevation on Lagoon History. — A gradual uplift of the sea-bottom, by decreasing the depth of water in front of the offshore bar, tends to cause the waves to break farther and farther seaward. If the elevation is so very slow that the normal tendency of the waves to deepen the water in front of the bar by erosion is not completely counteracted, the bar will retreat as on a stable coast, but more slowly. Should elevation just balance deepening by wave erosion, we should expect the bar to remain approximately in its original position while its crest was raised higher and higher out of the water and the lagoon became dry through emergence. Were elevation slightly more rapid, the waves would prograde the bar by adding successive ridges to its face as they broke farther and farther seaward. The older ridges would normally have a higher average crest elevation, through uplift, than would the later and hence less uplifted members of the series. The lagoon or marsh would disappear or dry up as the depression it occupied was raised above sealevel. Very rapid elevation might prevent the formation of well-developed ridges in front of the original bar; or if the rapid elevation began before any bar had formed, the bar and lagoon might not come into existence at all until elevation had ceased or become much more gradual.

Offshore Bars not an Evidence of Subsidence. — On an earlier page we have referred to the fact that certain authors are inclined to regard offshore bars as an evidence of coastal subsidence. We are now in a position to return to this theory, and consider it in the light of our discussion of the normal history of the offshore bar. It should be noted in the first place that in so far as the subsidence theory of bar formation has been elucidated by its supporters, it would seem to rest upon

one or the other of two misapprehensions regarding the history of offshore bars. McGee[32] and Ganong[33] assume that waves must begin to build up deposits of sand and gravel immediately at the margin of the original coast. Offshore bars must therefore represent former coast-margin beaches which have been built vertically upward as the land subsided and the receding shoreline moved inland. The assumption upon which the arguments of McGee and Ganong depend for their validity is, however, directly opposed to the conclusions of practically all students of shoreline phenomena, to theoretical considerations based on the principles of shoreline development as outlined above, and to observed facts.

It is not necessarily a serious objection to any theory to say that it is opposed to the conclusions of former investigators. Theoretical considerations, however, are directly in conflict with the assumption that offshore bars must have begun as ordinary shore beaches at the margin of the mainland. In our elaboration of the theory of shoreline development we have seen that the laws of wave action, according to which waves break in a depth of water about equal to the wave height, require a zone of breakers some distance from the mainland on a gently sloping shore. If waves breaking at the mainland margin erode the bottom and cast up part of the débris to form a beach ridge, we should expect larger waves breaking offshore to erode the bottom and cast up part of the débris to form an offshore ridge or bar. Moreover, according to the theory of wave action, subsidence, by deepening the water in front of the wave-built deposit, enables larger waves to attack the deposit in the effort to drive it landward. If waves could reach the mainland shore to build a beach deposit before subsidence began, it is difficult to see why more intense wave action under the more favorable conditions induced by subsidence should be unable to keep the deposit pushed back to the same relative position as the shoreline receded. The fact that the best development of offshore bars is found where geologically recent uplift has brought a smooth, gently sloping sea-bottom within the zone of effective wave action, is in accord with what we should expect if the theoretical considerations elaborated on preceding pages are correct; whereas the absence or poor development of such bars on many coasts known to have suffered subsidence in geo-

logically recent times is distinctly unfavorable to the theory which attributes such bars to subsidence.

Recorded observations prove that bars may be produced by waves breaking some distance out from the main shoreline. We have historical evidence of a few cases of this kind on a large scale, such as the Yarmouth bar on the east coast of England; on a smaller scale the process may be observed along the shallow shores of lakes and ponds. The writer has seen a very perfect miniature offshore bar formed in a few hours by waves raised on the surface of a small lake at Lakehurst, New Jersey, during a fresh breeze. The bar was a few inches in width, and separated a shallow lagoon one or two feet broad from the gently sloping sandy shore which it paralleled for some yards.

It is possible to read another meaning into the words used by McGee; and as this alternate interpretation may be held by others who regard offshore bars as proofs of subsidence, we will briefly consider it. In citing offshore bars (which he calls "keys") as an evidence of coastal depression, McGee uses the phrase: "the rapidly-encroaching sea having outstripped the slow-moving keys and left them far behind[34]." We might conceive this to mean that while the offshore bar was first formed by storm waves some distance out from the mainland shore, and possibly began to retreat landward under normal wave attack, subsidence intervened at so rapid a rate that the inner shoreline encroached upon the land faster than the bar could follow. Hence, one might argue, there is still a great breadth of lagoon or marsh between the bar and the mainland, whereas there would have been none by this time had it not been for subsidence.

The validity of this argument must depend upon two assumptions: first, that we know how long it normally takes a bar to move from its initial position to the mainland when *not* affected by subsidence; and second, that the bar was built that long ago. Neither of these assumptions is supported by any evidence thus far brought to light. We do not know the length of time required for an offshore bar on a stable coast to retreat to the mainland, nor do we know how long ago the bars on the New Jersey and other parts of our coast were formed We are not justified, therefore, in assuming that the persistence to

the present day of a lagoon or marsh back of the bar is in any wise related to coastal subsidence.

Mature Stage. — The offshore bar is a temporary feature, built by the waves because the initial slope of the upraised sea-bottom was not in harmony with the marine forces operating upon it. Once the bar is fully developed, and the steeper slope of its seaward side is brought into approximate adjustment with the waves which break against it, the normal retreat of the shoreline may begin. Constant loss of the finer products of attrition, which are swept into deep water by current action, enables the waves to drive the bar slowly landward. Temporary prograding may interrupt the retreat from time to time, as already explained; but such interruptions can have no effect on the ultimate history of the bar. It is inevitably forced farther and farther up the gentle slope of the lagoon bottom, or across the surface of the marsh deposits, toward the initial shoreline. The advancing waves cut farther and farther into the original sea-bottom in order to preserve the same depth of water immediately in front of the retreating bar. A time must come when the bar has been forced clear back upon the mainland, the lagoon or marsh has been wholly destroyed, and the steeper slope to deep water required by large storm waves lies just at the edge of the land. The shoreline of emergence is then said to be *mature* (stage *H*, Fig. 119).

Just as in the case of the shoreline of submergence, maturity of the shoreline of emergence is characterized by a very simple pattern. Indeed, it is apt to be much more nearly straight for long distances than is the mature shoreline of submergence, since it develops from a young shoreline which was itself straight or of simple curvature. The marine cliff bordering the shore may be very low and insignificant in early maturity, but will increase in altitude as the waves cut farther into the sloping coastal plain. When wave attack is vigorous the cliffs may themselves be young. This is especially apt to be the case during the early maturity of the shoreline. A narrow beach may intervene between the base of the cliff and the water; but owing to the changing profile of equilibrium under varying conditions of wave attack, the beach deposit may be temporarily removed and the bare rocky surface of the marine bench exposed for a time. In height the cliff will normally be more

uniform than that bordering a mature shoreline of submergence, since it is carved in the margin of a plain formed by the comparatively smooth uplifted sea bottom. The cliff line will be interrupted by the valleys of those main streams which are sufficiently active to cut their channels down to sealevel as rapidly as the waves push the shoreline inland. Smaller and weaker streams may descend from hanging valleys opening well up on the face of the cliff, the height of the valley mouth above sealevel being a measure of the relative incompetency of the stream which occupies it.

It is hardly probable that an offshore bar would retreat at such rate in all its parts as to reach the mainland shore simultaneously throughout its entire length. We must rather expect that a stage will occur when many parts of the bar touch the mainland, while along other parts narrow remnants of the lagoon still intervene between bar and inner shoreline. Especially will this be the case where the mainland shore was mildly irregular in outline. We may speak of such a shoreline as in the *submature* stage of its development, and cite the shore of the Landes district of southwestern France as an example. The cliffs at Long Branch on the New Jersey coast may be regarded as bordering a mature shoreline of emergence.

One effect of shoreline retrogression upon the drainage pattern of a coastal plain demands a word in this connection. Abbe[35] has described the asymmetrical position of the divides along the shores of Chesapeake Bay and its branches, where the water parting lies nearest to the shore which is retreating most rapidly. This is due in part, at least, to the fact that wave erosion cuts off the lower ends of the valleys faster than headward stream erosion can push the divide back to a position of stable equilibrium. The divide tends to migrate away from that shore which is retrograding most rapidly; but its migration is sluggish as compared with the rate at which the waves push the shoreline toward the divide. Hence the unsymmetrical position of the latter.

Old Stage. — There are striking differences between a young stream and a mature stream; but no such marked contrast exists between mature and old streams. Similarly, while the contrasts between young and mature shorelines are sufficiently remarkable to call for much comment, whether in the case of

shorelines of submergence or shorelines of emergence, little that is new can be said regarding old shorelines of either class. The remarkable lack of adjustment between shoreline and shore processes which characterizes the stage of youth, is replaced by a nearly perfect adjustment in maturity; and this same adjustment continues throughout old age. In like manner the relatively simple shoreline of maturity, bordered on the one side by a sufficient depth of water for wave action close to the land, and on the other by marine cliffs, persists into the latest stage of shoreline development. The water depth immediately adjacent to the shoreline may decrease as the marine bench is broadened and the movement of waves across it is retarded in old age; the cliff may weather back to a more gentle slope than it possessed during maturity, and hanging valleys may disappear as smaller streams become able to keep pace with the slower retreat of the shoreline. But such changes are of moderate importance as compared with the remarkable transformation which takes place between youth and maturity of the shoreline cycle.

It must not be forgotten, however, that the old age of a shoreline is largely a matter of theory. No good example of a shoreline in this stage of development is known to exist at the present time. Young and mature shorelines are well known; and from them we may reasonably infer what some of the characteristics of old age must be. On the other hand there are some questions concerning which we must speak with more reserve. Thus we have seen that as a land mass approaches the condition of peneplanation, the rivers can bring out very little débris, and waves will accordingly have less river-brought material to deal with. Under these conditions they may be able to attack more vigorously the task of eroding the coast and removing the wave-formed débris. How will the rate of shoreline retrogression then compare with the earlier rate? How will the depth of water near the shoreline, the slope of the marine cliff, and the condition of hanging valleys then compare with similar features at an earlier stage of the shoreline cycle? We might discuss such questions at length from the theoretical standpoint; but it would be difficult to confirm our theoretical conclusions by confronting them with facts observed in the field. Such consideration of these problems as appears to be profitable has already been given in Chapter V.

RÉSUMÉ

In the present chapter we have traced the development of the shoreline of emergence from the initial stage through its youth and maturity to old age. We have paused long enough to discuss at some length the origin of offshore bars, and have concluded that they are constructed for the most part of material eroded from the sea-bottom by onshore wave action, as was early stated by de Beaumont; although the action of longshore currents upon which Gilbert relied plays a significant rôle in their later history. It has been shown that the theory of tidal inlets which would explain their frequency and breadth as due to the amplitude of the tidal range, is in itself inadequate; and that the distribution of inlets, the breadth of lagoons, and the amount of lagoon filling are determined in part by the extent to which débris is transported along the face of the offshore bar by longshore current action. A study of migrating inlets has developed the important conclusion that an offshore bar broken by such inlets may exhibit the same cross section after migrating far landward over a salt marsh deposit as it did in its initial stage. We have given special attention to the effects of subsidence and elevation upon offshore bars, lagoons, and marshes; and have found that there is no support for the conception that offshore bars are an indication of coastal subsidence.

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CHAPTER VIII

DEVELOPMENT OF THE SHORELINE (Continued)

C. NEUTRAL AND COMPOUND SHORELINES

Neutral Shorelines. — It would not be appropriate in the present volume to discuss at length the developmental history of all the different types of neutral shorelines. The general principles outlined under the preceding discussion of shorelines of submergence and shorelines of emergence present a foundation upon which the student may base a treatment of any neutral shoreline, making such minor modification of treatment as the special peculiarities of the particular type may warrant. We may note in passing, however, that alluvial plain and outwash plain shorelines, like the shoreline of the coastal plain, have a simple pattern in the initial as well as in later stages; but that unlike the latter type, they need not pass through an offshore bar stage because of their steeper seaward slope from the water margin. Lobate delta shorelines pass through a submature stage in which an arcuate pattern is given to the outer shoreline by the building of bars connecting the seaward extremities of the lobes. Portions of the Rhone and Nile deltas appear to possess shorelines representing this stage of development. A true arcuate delta shoreline may characterize the mature stage of a lobate delta shoreline, if wave erosion cuts back the lobes beyond the heads of the inter-lobe bays (Fig. 120). An appreciation of the variety of delta types responsible for some of the variations in delta shorelines may be gained from an inspection of Credner's well-known essay on "Die Deltas[1]".

A valuable discussion of delta formation is given by Barrell in an essay on " Criteria for the Recognition of Ancient Delta Deposits." The "delta cycle" is thus briefly summarized by this author: " In the stage of youth before the drainage system has become well developed the detritus delivered at the river mouth is somewhat smaller in amount but coarser in texture. The subaqueous wave-cut profile is also undeveloped, the bottom still inheriting

its original slope. If this initial slope is gentler than the subaqueous profile of equilibrium the waves have at first less power of erosion at the coast line. If the initial slope is steeper they will possess an initially greater power. Assuming, however, that the river is dominant over the sea, the delta is rapidly

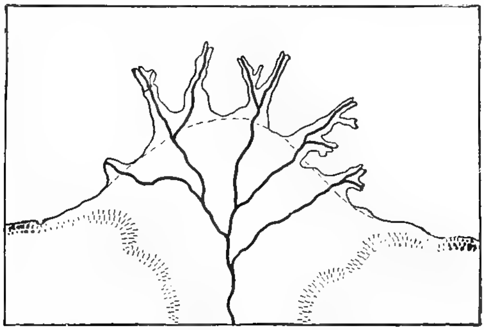

Fig. 120. — Diagram showing how wave erosion of a lobate delta may transform it into an arcuate delta (broken line).

built outward, and on account of the coarse waste, the steeper river grades, and shallow bottom near shore, the initial proportion of the subaërial topset beds is relatively high. During maturity the quantity of waste is larger, as all parts of the drainage system now supply sediment, but as the river is graded and its gradient is also flattened the waste is finer in texture. The delta is extended outward and the greater deposit is on the outer portions. It grows inland also for a time, but owing to the flattening grade the beds in this direction show decreasing thickness. The maximum *rate* of outward growth is reached early because of the increasing surface area, which requires a greater volume of sediment to give a unit thickness, and the increasing depth of the water, which involves a continually deeper fill. Furthermore, the increasing shoreline and greater exposure to the waves increase the power of the latter to carry away the waste, which with the progress of the cycle becomes finer in texture and more readily removed by the sea. But although the rate of advance falls off, the outward growth will continue during the progress of maturity in the cycle of erosion and deposition. In old age, however, on account of the ever-slackening supply of waste and the larger portion carried in

suspension and solution, the sea at last gains the mastery and begins to plane inland across the low-lying and unconsolidated materials projecting into the sea. Rapid headway is finally made against the weakened river; the territory conquered by the river in its youth is reclaimed and the sea at last will beat once more against the margin of the old land[2]".

The development of fault shorelines has been ably discussed by Cotton[3], who presents a detailed analysis of the features

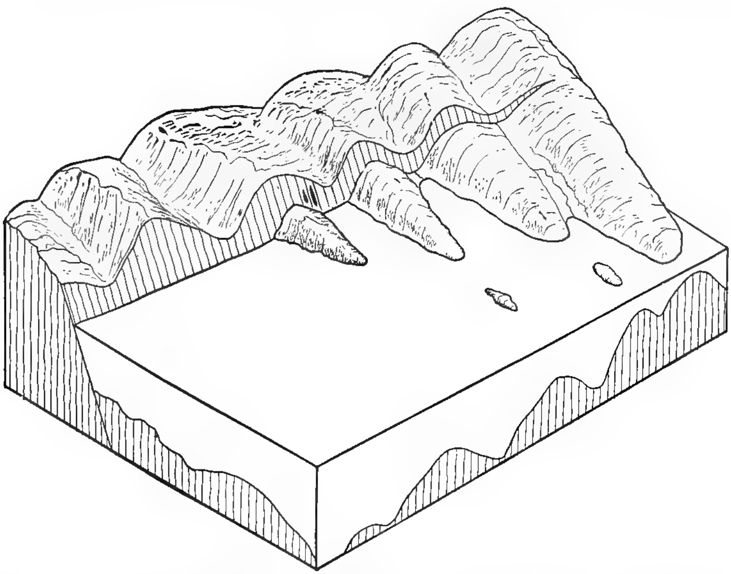

Fig. 121. — Fault shoreline bordering a scarp which dies out toward the right. The fault traversed a region of strong relief. (Modified after Cotton.)

characterizing successive stages in the life history of such shorelines. As he is careful to point out, the initial character of the fault shoreline will vary widely according as the fault traverses a maturely dissected land mass of strong relief (Fig. 121), or an undissected coastal plain of no appreciable relief (Fig. 122). In either case streams betrunked by faulting will cascade into the sea from the mouths of hanging valleys. Thus the initial stage of fault shorelines resembles the mature stage of shorelines of submergence in cases where the relatively simple cliffline of the latter type is marked by hanging valleys due to

rapid wave attack. The character of the seaward slope, however, is very different in the two cases. Where the landward block bordering a fault shoreline has itself been partially depressed, thereby bringing the main valley floors at the fault scarp down to sealevel, there will be no large hanging valleys.

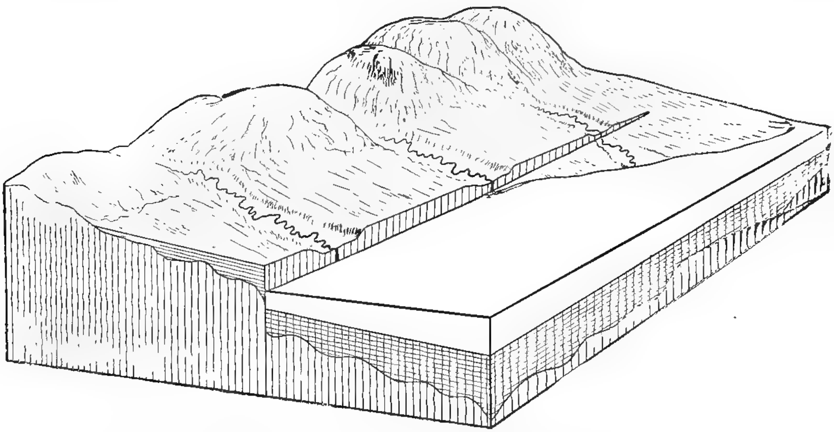

FIG. 122. — Similar to Fig. 121, except that the fault traversed a little-dissected plain of faint relief. (Modified after Cotton.)

If the landward block has been depressed sufficiently to permit the sea to enter and submerge the main valleys, we have an initial compound shoreline (Fig. 124), the treatment of which type is reserved for a later section. Seaward from the fault scarp the sea floor will have the contours of the pre-faulting land surface, whether that be an irregular surface (Fig. 121) or a smooth plain (Fig. 122). The seaward slope in the immediate vicinity of the shoreline will normally be very steep, as it is the slope of the fault scarp itself.

Wave attack on the fault scarp will not proceed very rapidly at first, both because steep walls rising out of deep water tend to reflect waves, and because the waves are unarmed with rock fragments with which to make their attack more effective. The face of the cliff will weather back to a more moderate slope, and the weathering products will accumulate at the cliff base as a subaqueous talus. Streams emptying from hanging valleys will rapidly entrench themselves, cutting young gorges in the more mature valleys of the initial land surface, thus producing a

typical two-cycle topography without necessarily implying any change in the level of the land area in question or of the adjacent water surface. The erosion products brought out by the streams will accumulate as subaqueous talus cones which may later take the form of ordinary deltas. With the shallowing of the bottom near shore by accumulations of débris derived from fault face and stream valleys, wave reflection is less perfect and wave attack more vigorous, particularly since supplies of

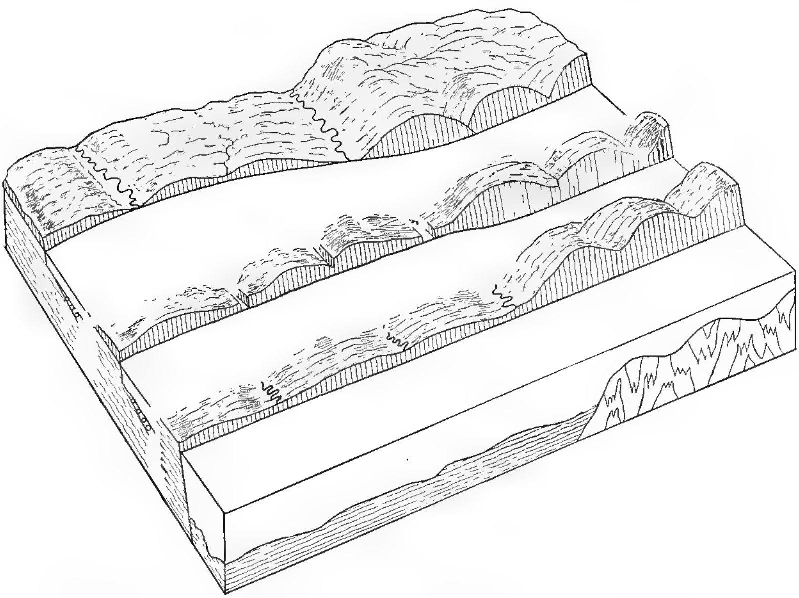

FIG. 123. — Successive stages in the retrogression of a fault shoreline bordering rocks of varying resistance.

rock fragments are now accessible to the waves. The retreat of the shoreline takes place more rapidly for a time. Later, when the marine bench and shoreface terrace have attained a considerable width, the vigor of the waves traversing them is somewhat reduced for reasons explained on earlier pages; and the marine cliff, steepened while wave attack was increasing in vigor, has opportunity to weather back to a more gentle slope. The fault shoreline has now reached maturity, and henceforth develops in the same manner as other mature shorelines. As noted under shorelines of submergence, a shoreline bordering

weak rock areas will retreat more rapidly than one bordering regions of resistant rock. From this it follows that an initially straight fault shoreline may acquire a pattern of simple curves in maturity, the re-entrant curves being systematically related to weak rock areas (Fig. 123).

Compound Shorelines. — Thus far we have considered the developmental stages of shorelines of submergence, shorelines of emergence, and neutral shorelines. It remains only to point out very briefly any special features characteristic of different stages of compound shorelines, or those shorelines which exhibit prominently features normally characteristic of at least two of the foregoing classes.

In its young stage a compound shoreline combining features of both submergence and emergence will be characterized by an offshore bar which determines a straight outer shoreline, and drowned valleys which give an irregular inner shoreline. The bar may be broken by tidal inlets and possess all the other features of such a bar on a shoreline of emergence. Similarly, the lagoon may in course of time become filled with sediment or marsh deposits. Whether or not such filling occurs, the various types of spits, forelands, and bars which are so marked a feature of young and submature shorelines of submergence, are largely lacking along the irregular inner part of the compound shoreline, for the reason that the offshore bar protects the inner shore from the effective wave action which is, as we have already seen, largely responsible for these shore forms.

Such a compound shoreline may be called submature when the offshore bar has been driven against the headlands of the inner shore. The bays between the headlands will then appear to be closed by bay bars; and cases may occur in which a submature compound shoreline could not be distinguished from a submature shoreline of submergence. In the latter type of shoreline, however, the bars closing the different bays have developed more or less independently; and it is doubtful whether they will ordinarily show that relatively straight alignment characteristic of the different parts of a single offshore bar which has been driven against the headlands of the irregular inner part of a compound shoreline. Maturity is reached when outer and inner shoreline have coalesced in one shoreline back of the heads of the initial embayments. From this time on the features of

the compound shoreline do not differ from those of the shoreline of emergence.

A compound shoreline combining features of a fault shoreline with those of a shoreline of submergence (Fig. 124) passes through a first stage in which the outer or fault shoreline portion develops like any normal fault shoreline, while the drowned valley portions of the partially submerged block have the same history as the more deeply indented portions of normal shore-

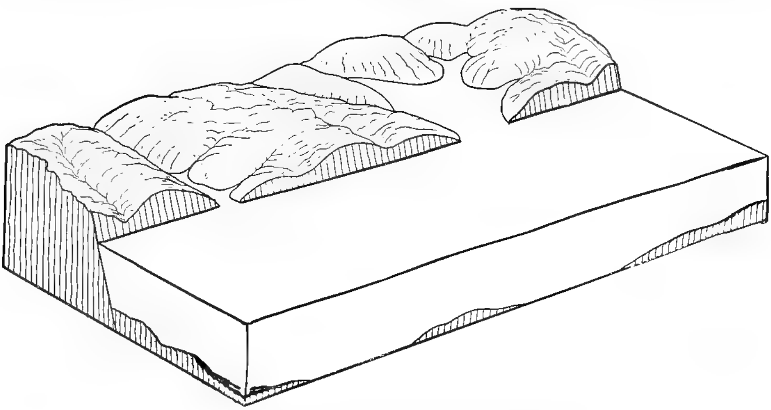

Fig. 124. — Compound shoreline, combining essential features of a shoreline of submergence and a fault shoreline.

line of submergence. Maturity is reached when wave erosion has pushed the initial fault scarp, later become a normal marine cliff, back of the bay heads, and a simple shoreline bordered by a continuous marine cliff is developed. Further stages of development show no peculiar features.

Contraposed Shorelines. — If a coastal region of hard rocks is separated from the sea by a belt of overlapping softer deposits, as where a rugged oldland is overlapped by a narrow coastal plain, the shoreline which is first developed upon the softer beds will later be retrograded until it comes against the hard rocks. Such a shoreline has well been called "contraposed" by C. H. Clapp[4], and in origin it is analogous to a "superposed" river which has been let down from a soft rock cover upon underlying ridges of harder material. A shoreline which has reached maturity in the softer beds may in its contraposed position lose

PLATE XLVI.

Contraposed shoreline south of Rye Beach, New Hampshire. The waves have cut back the original shore developed in glacial drift until they have discovered resistant crystallines.

its mature characteristics and acquire those of youth (Fig. 125). It may even change from a typical shoreline of emergence to one having the characteristics of submergence, if the older and

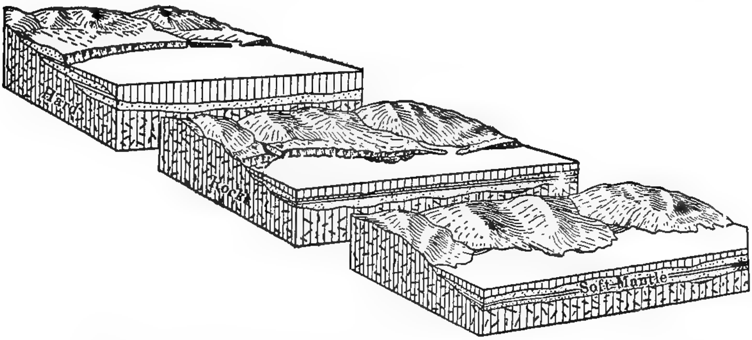

Fig. 125. — Stages in the formation of a contraposed shoreline. Early stage shown by upper figure. (Modified after Clapp.)

harder rocks possessed a very rugged surface and the soft rock mantle consisted of unconsolidated material easily removed. Parts of the New England shoreline belong to the contraposed type (Plate XLVI).

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CHAPTER IX

SHORE RIDGES AND THEIR SIGNIFICANCE

Advance Summary. — Many beaches, bars, tombolos and forelands are characterized by a succession of narrow ridges built by the waves, and sometimes later modified by the winds. These "lines of growth" of shore forms have much significance for the engineer who would learn something of current action and direction of débris movement at a given locality in the recent past, and for the geological or geographical student who would trace the development of shore forms and ascertain what light they may throw upon the important question of past changes in the relative levels of land and sea. It is the purpose of the present chapter to discuss the origin of beach ridges and dune ridges; to inquire into the rate at which they have been formed, with the hope of acquiring data useful in estimating the ages of those shore forms which possess them; and, finally, to analyze in a critical manner the conditions under which beach ridges may be used to determine whether coasts have recently experienced appreciable changes of level.

Origin of Beach Ridges. — Beach ridges have long been recognized as representing successive positions of an advancing shoreline, and are known to the English as "fulls"; while the depressions between them are known as "swales," "slashes," or "furrows." When a beach ridge is covered by dune sands we have a "dune ridge"; the swales between dune ridges have been called "dune valleys" (Dünentäler) by the Germans. Unusually good examples of beach ridges or dune ridges are found on Orford Ness on the east coast of England as described by Redman[1]; on the Dungeness foreland of the southeast coast, described by Drew[2], Redman[3], and Gulliver[4]; on the Darss foreland of the Baltic coast of Germany described by Otto[5] in a paper on "Der Darss and Zingst"; at Swinemünde on the same coast where a most remarkable series of dune ridges has been described by Keilhack[6] in a most interesting essay entitled "Die

Verlandung der Swinepforte"; and on Cape Canaveral off the east coast of Florida. Beach or dune ridges are usually found in a greater or less degree of perfection on any prograded shoreline, and they are of so much importance, not only in showing the successive stages of development of the forms which possess them, but also, as will presently appear, in showing whether a coast has remained stable or experienced changes of level during their formation, that it is pertinent to inquire somewhat fully into their origin and significance.

According to Gilbert[7] a wave-built terrace or beach plain is usually produced whenever a shoreline maintains its course while the longshore current diverges. "The surface of the wave-built terrace, considered as a whole, is level, but in detail it is uneven, consisting of parallel ridges, usually curved. Each of these is referable to some exceptional storm, the waves of which threw the shore drift to an unusual height[8]". The forward building of the shore occurs because the diverging current assumes a greater cross section and a diminished velocity; and with diminished velocity an accumulation of the transported débris must take place. "This accumulation occurs, not at the end of the beach, but on its face, carrying its entire profile lakeward and producing by the expansion of its crest a tract of new-made land."

Davis[9] explains the prograding of an offshore bar by supposing that waves breaking on a shallowing sea floor cast up the bottom material into an initial bar or ridge; later, larger storm waves break a little farther out in deeper water, and from the newly eroded bottom material construct another bar on the face of the earlier one. "A preliminary offshore bar is built up by the storm waves . . . ; and afterwards, at times of exceptional storms, successive additions may be made on its outer side[10]." According to this theory, also, the accumulation occurs on the face and not on the end of the earlier deposit; but the material is supposed to be derived from the sea-bottom and not from the longshore currents upon which Gilbert relied.

When a recurved spit develops into a compound spit or foreland by the addition of successive spits or embankments to its seaward side, there is produced a beach plain characterized by sub-parallel ridges separated by belts of lower land or strips of water. In this case, however, the accumulation may take place

PLATE XLVII.

Ancient beach ridge (center of view) connecting former island in distance with one in foreground. A later beach ridge to the left. Note that a road has been built on the crest of each ridge, and that the intervening swale is spanned by foot bridges.

not simultaneously along the entire face of the earlier deposit, but by extension of the ends of the successively formed embankments; longshore currents furnish the principal supply of material; and the individual ridges are evidently not to be correlated with a corresponding number of great storms. Davis[11] appears to regard the beach ridges of the Provincelands as having been produced in the manner here indicated, although his admirable essay on "The Outline of Cape Cod" does not explicitly state that the successive embankments all grew longitudinally from their point of tangency with the mainland cliff.

It is highly probable that ridged beach plains have been produced in all three of the ways mentioned above. Where one part of a shore is being cut back and straightened by the waves, a longshore current may have its course so modified as to depart from an adjacent section of the shore which it previously followed. If the withdrawal is gradual enough, the portion of the shore affected may continuously be prograded by deposits laid down in the manner described by Gilbert. Where the withdrawal is more rapid, successive separate embankments may build a compound spit or foreland bar. Intermediate forms between these two types must exist. Sandy Hook in its earlier development appears to have consisted of several embankments built independently from southeast to northwest, one after the other; but in later years it seems possible that its whole seaward face has advanced eastward at times by practically simultaneous deposition along its length. There is likewise good reason to believe that some offshore bars have been slightly prograded by the building of one or more embankments in the deeper water outside of the original bar, after the manner suggested by Davis. In the opinion of the present writer, however, the processes described above are not the only ones, nor perhaps the most important ones, by which ridged beach plains are produced; nor should the beach ridges in any case be regarded as the product of individual great storms, as has been so commonly assumed.

It has already been shown in connection with the discussion of beach profiles of equilibrium, that a shoreline must be prograded wherever longshore currents of any type bring to it more débris than the waves there operating can remove. Deposition of excess débris shallows the offshore bottom, favor-

ing the formation of waves of translation, which in turn drive the bottom débris on shore until prograding of the shoreline and deepening of the bottom produces a profile which is in equilibrium with the forces there at work. If the supply of débris by longshore currents is kept up indefinitely, the shore may be extensively prograded before equilibrium is established. It should be noted that in the case here considered the longshore currents are forced to move seaward because the shore is prograded, whereas in the case mentioned by Gilbert the shore was prograded because the currents moved seaward. Waves are the active agents in causing the prograding, and derive much of their material from the offshore bottom, as in the case of offshore bars mentioned by Davis. But unlike the case considered by him, longshore currents are primarily responsible for a continuous supply of material which as continuously shallows the offshore bottom; and the prograding of the shore is not to be correlated with the initial bottom slope nor with storm waves of different sizes. The shoreline advances seaward throughout a considerable portion of its extent simultaneously, and does not grow by the longitudinal extension of each ridge, as in the case of compound spits.

It is immaterial what particular type or types of currents bring an excess of débris to the prograding area. Beach drifting along both the shore and shoreface zones is an exceedingly important and commonly neglected source of supply. Where beach drifting is from opposite directions toward a common point, as not infrequently happens in bays and lakes, there will be an accumulation of material at the meeting point, where weaker or conflicting wave currents are unable to dispose of it. Beach drifting in but one direction along a shoreline which suddenly changes its trend, will cause an excessive deposit just beyond the angle in case the shore bends backward, because wave action upon the more protected shore around the bend is not sufficiently vigorous to remove all the débris deposited there Material drifted along bayside beaches toward the bay heads, shallows the latter areas and permits the small waves operating there actively to prograde the shoreline.

Offshore bars are characteristic of shorelines of emergence, and are described on an earlier page. But it will not be inappropriate to consider in this connection the origin of pro-

graded bars showing beach ridges. Where an offshore bar has been formed with a profile of equilibrium nicely adjusted to the marine forces along its entire length, it is evident that any disturbance of conditions at one point along its sea front may lead to retrograding or prograding at another. A succession of storms causing unusual erosion in one locality may permit beach drifting or other longshore movements to carry an excessive amount of débris to another part of the bar, disturbing the equilibrium there and causing prograding. The opening and closing of inlets, by affecting longshore transportation, may indirectly cause retrograding or prograding on adjacent parts of the shore. Additions to the face of an offshore bar do not necessarily imply, therefore, that larger storm waves have been breaking on the deeper parts of an initial sloping sea-bottom; neither does retrograding indicate that the bar previously advanced to the zone where the largest storm waves broke on the initial bottom, and that it has now entered a new stage of its development characterized by progressive retreat. On the contrary, both retrograding and prograding must frequently be interpreted as horizontal oscillations of the shoreline consequent upon disturbances of the shore profile of equilibrium which may be very temporary in some cases, but endure for a considerable time in others. As will be shown in later chapters, parts of the Atlantic shoreline have repeatedly been retrograded and prograded. It follows from these considerations that the retrograding or prograding of a shore does not form a satisfactory basis for discriminating between stages of shoreline development, as has been sometimes assumed.

It may happen that an initial shallow on a shoreline of submergence will for a long time occasion the formation of waves of translation, which will in turn sweep upon the shore all débris deposited over the shallow. A cuspate foreland may thus advance over the shallow and finally conceal it, with the result that the shore will exhibit a foreland unrelated to any visible shore irregularity or any known currents. A river may deposit so much sediment opposite its mouth as to shallow the sea-bottom, whereupon the waves will re-establish the shore profile of equilibrium by eroding the bottom and prograding the shoreline, the latter action producing a cuspate foreland (or cuspate delta) showing parallel beach ridges (Fig. 126).

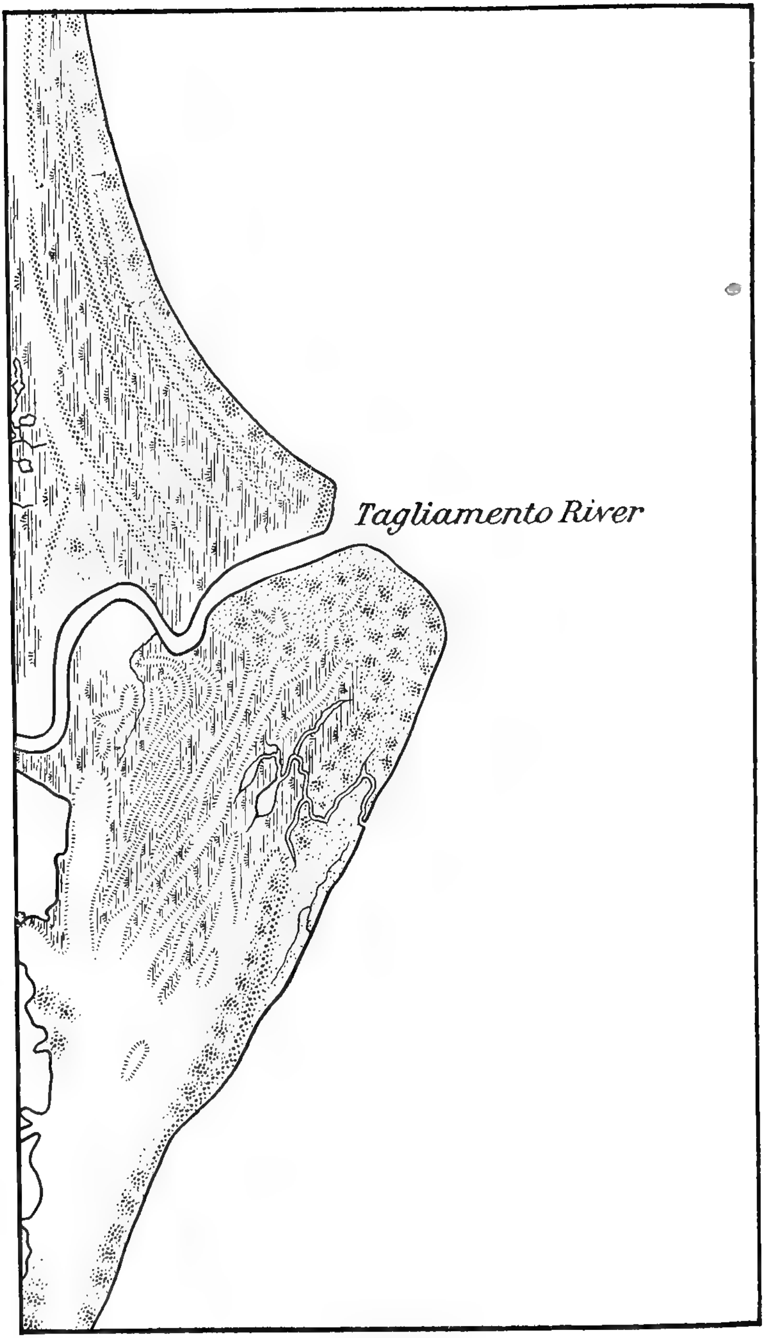


Fig. 126. — Cuspate delta of the Tagliamento River, Italy, showing parallel beach ridges.

There is a widespread belief that the beach ridges, which often characterize the surface of forelands, bars, tombolos, and other prograded shore forms, represent the work of individual great storms. Men whose opinions must always carry great weight have either explicitly or by implication supported this view. Gilbert[12] is very clear in his statement: " Each of these (ridges) is referable to some exceptional storm, the waves of which threw the shore drift to an unusual height." Davis[13] expresses much the same opinion concerning beach ridges on offshore bars, but adds that further study and observation are required to demonstrate the validity of certain points in his explanation of bar formation. Other authors have expressed somewhat different views. In discussing a paper by Redman[14] on the shore deposits along the south coast of England, B. S. Howlett[15] states that every beach ridge represents " the accumulation of shingle resulting from some stormy tide," while Sir William Cubitt[16] " apprehended that these 'fulls' coincided with, or at least were influenced to some extent, by the lunar cycles." Cornish[17] would recognize " neap tide fulls " and " spring tide fulls." He apparently considers that these tidal ridges may be amalgamated into a " summer full " and a " winter full," and that these larger fulls may in their turn sometimes coalesce. Unlike most observers, Wheeler[18] believes that the ridges were built up during calm weather. Solger[19] advances the theory that in the case of dune ridges, which as we shall see later are essentially beach ridges capped by sand dunes, each ridge was formed during a dry climatic period, when the sand of a prograding shore was blown back to the line of ridge formation; while the intervening swales represented wet periods during which vegetation advanced rapidly over the newly gained land and prevented the sand from being blown into dunes. Three dune ridges are supposed by Solger to be formed each century, each of which corresponds to the dry phase of the well-known 35-year climatic period of Brückner. Keilhack[20] estimates that at Swinepforte one ridge has formed in every 35 years on the average, and he follows Solger in correlating their formation with the 35-year Brückner cycle.

There are several reasons for doubting the possibility of correlating individual beach ridges with a corresponding number of exceptional storms which cast up the shore drift to an unusual

height. In the first place, it is difficult to imagine the supply of shore débris and other shore conditions so adjusted that each exceptional storm would find enough material available with which to construct a high ridge, yet too much to permit the ridge to be driven back into coalescence with an earlier one formed by the last preceding exceptional storm. On the contrary, we should rather expect that one exceptional storm might do no more than raise a submarine bar in front of the shore; a second great storm from a slightly different direction might wipe the bar out of existence; the bar might reform during a third storm of equal violence; moderate waves in calmer weather might then raise the surface of the bar into a ridge a number of feet above sealevel; the next great storm might produce a new bar in front of the one just formed; and so on. In this imaginary case there occurred four exceptional storms, but there are only two beach ridges; and one of these was not raised above the sea by any of the storms. Observation will show that many beach ridges when followed along their crests subdivide into two or more ridges. Manifestly, if the separate ridges be regarded as the work of several exceptional storms, the compound ridge cannot properly be regarded as the work of one storm. The number of ridges formed in a given time do not correspond with the expectable number of great storms within that period. Thus the 121 ridges of the Darss foreland in Germany have been built in a period estimated to be from 3000 to 6000 years which would mean an average of only one great storm in every 25 or 50 years. If the time required for the development of Nantasket beach has been correctly estimated by Johnson and Reed[21] one would have to suppose that only one great storm in several centuries has been recorded by the beach ridges in the southern half of that district.

It is clearly impossible to suppose that every great storm builds a beach ridge, for observation abundantly proves the contrary. Indeed, I know of no case in which a typical complete beach ridge of large size has been wholly produced by one storm, although I do not regard this as impossible. On the other hand, a large part, if not all, of a beach ridge is often swept away during a single exceptional storm. We cannot suppose that every beach ridge represents the work of one exceptional storm, since, as has been shown, such a ridge often

represents the combination of several ridges elsewhere distinct, I do not believe that one should even regard a given beach ridge as necessarily the product of several exceptional storms; for while unusually high beach ridges must have been subjected to the influence of waves of sufficient magnitude to cast débris to their crests, the majority of ridges could have reached their present height through the influence of ordinary storm waves, and many of them perhaps by very moderate wave action at high tide. It is even possible to suppose that on a given beach plain none of the exceptional storms of the past are recorded by any of the ridge crests, but only the more prolonged activities of less violent wave action.

The height of a beach ridge depends in part upon the size of waves, but in part also upon other factors, among which may be mentioned rapidity of supply of material, and the relation of the new ridge to pre-existing ridges. If longshore currents supply débris with great rapidity, the shoreline may be prograded so fast that a given beach ridge has little opportunity to grow to a great height before the shoreface zone is shallowed and a new ridge begins to form in front of it. A number of ridges of moderate height might thus be formed in the intervals between exceptional storms. Continued shallowing of the offshore zone due to rapid deposition would also tend to change the largest storm waves into smaller waves of translation before they reached the line of ridge building, with the result that even great storm waves might not build high ridges. Less rapid supply of shore débris would favor the building of higher ridges in several ways: waves could cast material upon the ridge nearly as fast as it was supplied, enabling a ridge to grow to its full height before sufficient change occurred in the shore profile to require the initiation of a new ridge farther seaward; great storm waves would have a better opportunity to reach the shoreline, and the longer life of a ridge at the shoreline would increase the chances of such waves assisting in its construction; and while slower débris supply would increase the danger of ridge removal by storm waves, it would also increase the chances that wave attack might drive the shoreward ridge back upon the one behind it, thus forming a compound ridge of greater height. It can hardly be doubted that many of the prominent beach ridges of prograded shores represent the accu-

mulations of many subordinate beach ridges successively formed in front of a main shore ridge and later driven back upon it.

The future of any given beach ridge is very uncertain, because of the variable nature of the marine forces operating upon a prograding shore. It may have its further growth arrested by the development of another ridge in front of it; it may be completely washed away by the next storm; it may grow until it acquires large size and permanence of position; or it may be driven back to coalesce with one or more earlier ridges. A ridged beach plain is thus a very imperfect record of a complex history: only a fraction of the ridges once formed are preserved; the records of many storms are forever lost; some of the remaining ridges may record one great storm, others certainly represent the work of many different wave attacks upon the same line, while still others are composed of two or more formerly independent ridges forced into coalescence. One may admit that beach ridges can be materially affected by great storms, by spring and neap tides, by summer and winter storms, and possibly even by a 35-year climatic cycle; but he must still recognize the impracticability of correlating a given series of ridges with a given succession of any of these phenomena.

Rate of Beach Ridge Formation. — The student of shorelines often desires to secure an approximate idea of the length of time which has elapsed since the sea worked upon a certain part of the coast, and a succession of beach ridges sometimes affords the best available data. It is occasionally possible to determine the time occupied in building a certain number of the latest ridges, and if the rate were uniform throughout the growth of the entire beach plain the problem would be a simple one. From what has been said, however, it is evidently far from safe to assume that the older ridges were formed at the same rate as those of later date. The history of a beach plain is too complex, and its record preserved in too incomplete a manner, to enable one to say how few or how many ridges have been eliminated by erosion or coalescence. Furthermore, the rate of débris supply must vary with time, and the increasing depth of water encountered as the plain builds forward into the sea must affect its rate of growth. There are, nevertheless, certain general principles which may guide one in endeavoring to reach a reasonable conclusion as to the approximate time

Ancient series of dune ridges on Cape Canaveral, truncated at right angles by a later series in the foreground, on one member of which the man is sitting.

represented by a given series of ridges; these may be stated categorically, with such comments as seem necessary.

1. Short ridges normally require less time than longer ones. Thus a series of short ridges representing successive recurved points at the end of a spit may succeed each other with rapidity, since all the débris carried along the shore is concentrated at the narrow end of the spit. The same amount of débris employed in prograding a long stretch of the shoreline would build very few long ridges in the same length of time. Rockaway Beach, near the entrance to New York Harbor, is a good example of a compound recurved spit which is growing westward at a fairly rapid rate by the addition of successive recurved ridges of small height at its distal point. As will appear from Figure 127, reproduced from survey charts and reduced to a common scale by the Metropolitan Sewerage Commission of New York City, the westernmost ridge of the 1889 chart was diminished in area and two new ridges were added before the survey of 1905. Seven years later three additional ridges had been formed. In other words, five ridges were formed in twenty-three years, the average rate of growth varying from one ridge in eight years to one ridge in a little more than two years. The actual increase in the length of the spit during the whole period was nearly one mile, or an average annual advance of over 200 feet.

2. For a given exposure, low and narrow ridges imply a smaller lapse of time than an equal number of high, broad ridges. This depends upon the fact already explained that rapid supply of débris tends to cause a rapid prograding of the shoreline, with opportunity for low and narrow ridges only to form.

3. One series of parallel ridges abruptly truncated by another series trending in a different direction (Plate XLVIII), does not necessarily imply a longer lapse of time than would a single parallel series containing the same total number of ridges. This will be apparent from Figure 128. Let us imagine that a projecting headland, with the shoreline 0, 0, is cut back on its north side to the new shoreline 1, and that the eroded débris is deposited on the east to form the beach ridge 1′. Later erosion cuts the shore farther back to 2, thereby removing the extreme northern end of the beach ridge 1′, while deposition of the eroded débris forms beach

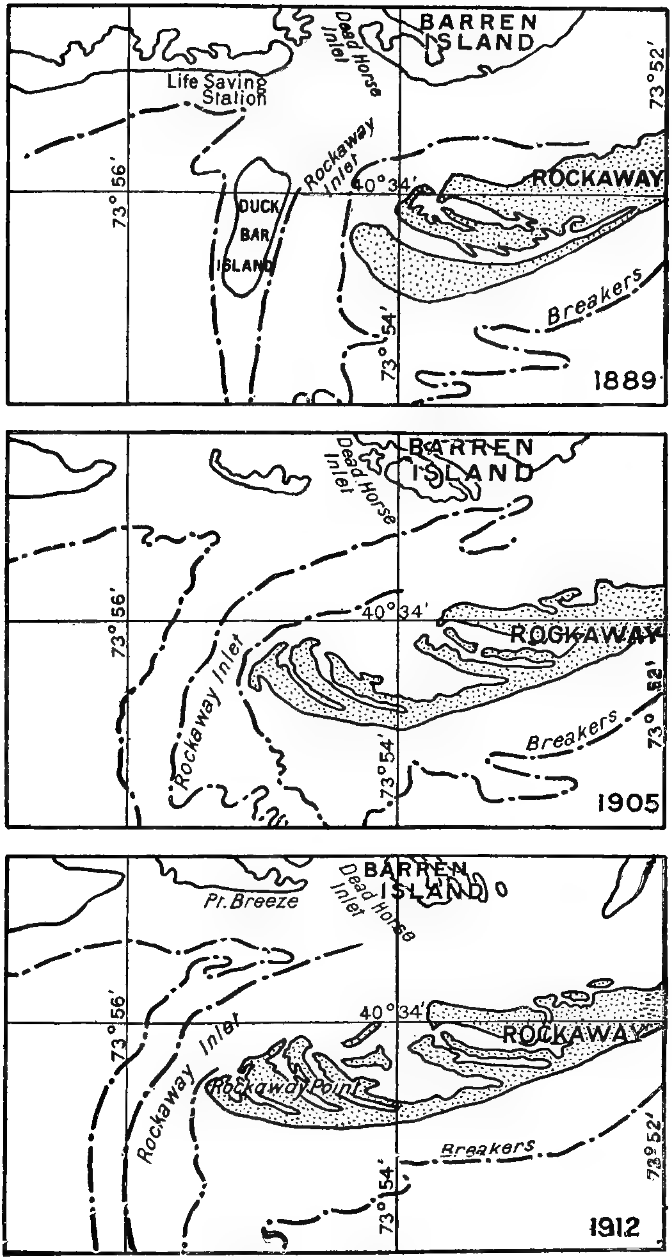


FIG. 127. — Successive stages in the development of Rockaway sand spit, Long Island.

ridge 2′. This process is repeated, until erosion drives back the shore to 5, thereby truncating the northern ends of beach ridges numbers 1′ to 4′ inclusive, and deposition forms beach ridge 5′. Owing to a change in the balance of shore processes, possibly consequent upon a change outside of the area shown in the figure, deposition replaces erosion on the north, and the beach

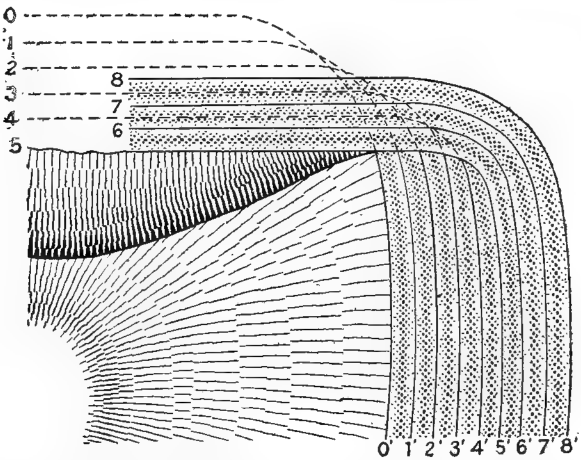


FIG. 128. — Diagram of cliffed headland and associated beach ridge plain, showing that one series of ridges truncating another does not necessarily imply a longer lapse of time than an equal number of parallel ridges.

ridge 6, 6′ is formed all around the headland, being followed by ridges 7, 7′ and 8, 8′. We now have on the north one series of parallel ridges which abruptly truncates another series; yet no greater time is here represented than that represented by the continuously parallel series 1′ to 8′ measured toward the east. It is not safe to assume, as has sometimes been done, that where one series of ridges truncates another, allowance must be made for a large time interval at the break. Discordance of ridge direction may or may not imply a greater lapse of time than accordance.

4. Dune ridges, or parallel ridges of dune sand corresponding in all respects with beach ridges, except as regards details of surface form, are to be regarded as resting upon true beach ridges, and may be used as readily as the latter in interpreting shoreline

changes. The regularity of crestlines and parallel arrangement of the dune ridges in such regions as the Darss foreland and Cape Canaveral (Fig. 129) leave no doubt that they are lines of shore dunes which have formed on the successive beach ridges when each ridge was next to the sea. Trenches cut through dune ridges have revealed the presence of beach sands or gravels below. That the dune ridges have not moved from their initial position is evident, for had they done so their crests would have become very irregular and would necessarily have lost their beautiful parallelism. On Cape Cod, where the dunes of the Provincelands have migrated under the influence of the winds, their former parallelism is lost and the position of the beach ridges is scarcely determinable. The idea, sometimes advanced, that the ridges are merely lines of shore dunes which have rolled inland from a stationary shoreline like waves of the sea, will not commend itself to those familiar with the phenomena of dune migration. Since the dunes must have formed in place on beach ridges at the shore, there must have been time enough in each case for a beach ridge to be formed by the waves; probably also for enough vegetation to gain a foothold on the ridge to arrest windblown sand coming from the beach and so prevent its being carried over into the swale or slash back of the ridge; and, finally, for the dune sand to accumulate in sufficient quantity materially to augment the height of the ridge. Long, broad, and high dune ridges, like those of the Darss or Canaveral, must have required many years for their construction.

5. While a large number of beach ridges indicates the lapse of a long time interval since the first one was formed, the converse is not true. On a graded shoreline, where neither prograding nor retrograding is occurring, a single beach ridge may represent the slow accumulation of many centuries. Not infrequently two or three beach ridges on one part of a shore represent the same time interval as does a large series of beach ridges on a closely adjacent part of the shore. It is not permissible, therefore, to assume a short time interval for the building of narrow beach plains containing but few ridges.

Dungeness Cuspate Foreland.—It may not be without interest to review some of the data available as to actual or estimated rates of beach ridge and dune ridge formation. The Dungeness of southeastern England is a prominent cuspate foreland project-

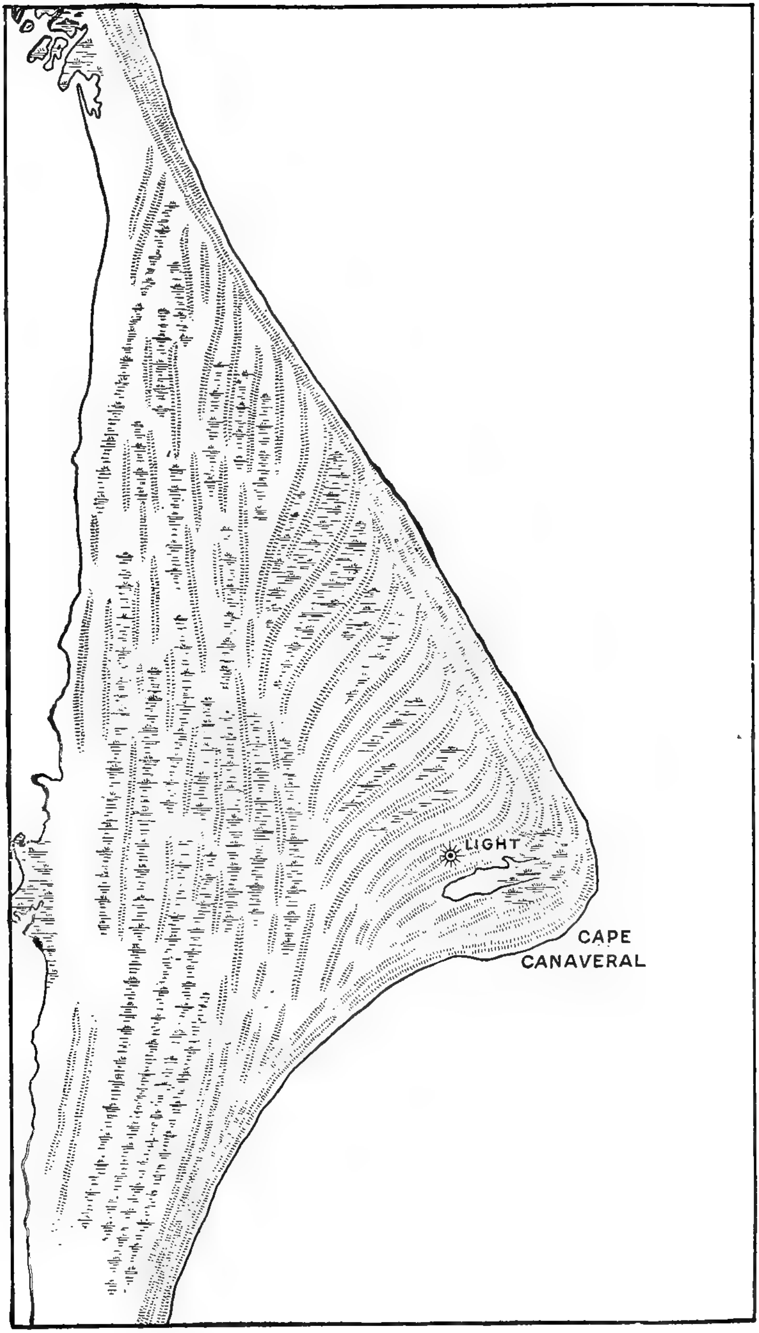
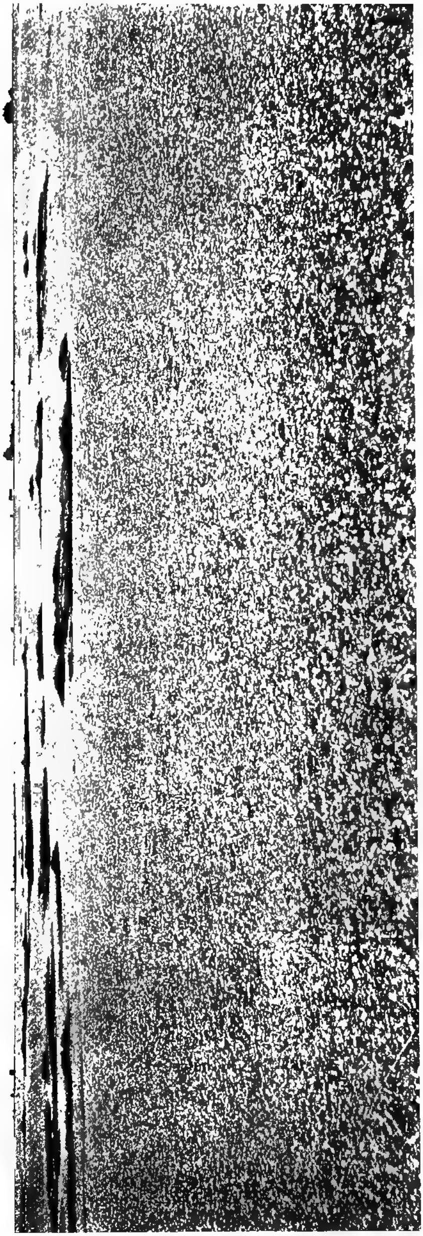


FIG. 129. — Ridges of the Cape Canaveral cuspate foreland. For the most part they are dune ridges, but beach ridges little altered by wind action occur near the Light.

PLATE XLIX.



Shingle beach ridges of the Dungeness cuspate foreland, England.

ing from a curved reëntrant of the shore and measuring about 15 miles along either side, its seaward portion consisting of a splendid series of shingle beach ridges. Between certain groups of the ridges are broad belts of marsh, while the base of the foreland consists almost wholly of marshland formed by the silting up of an extensive bay, which formerly occupied the interior of an initial compound cuspate bar. As a rule the shingle ridges are covered by very little vegetation, although some of the older ones are grassed over; while belts of grass and broom occupy many of the swales, thereby emphasizing to the eye the ridged character of the surface. (Plate XLIX.) In general appearance the ridges and swales closely resemble the well known "Wällen" and "Rinnen" of the island of Rügen (Plate L) described and figured by Braun[22]. So far as I could judge without careful measurements, the ridge crests of the Dungeness are prevailingly of moderate height, possibly rising 3 to 6 feet above intervening swales and 8 to 12 feet above high tide, where typically developed. Occasional sandy ridges are encountered, but well rounded flint shingle is the only material found in most of the ridges.

As shown by the map (Fig. 130) the older ridges have clearly been truncated by wave erosion on the south side of the foreland or "ness," and the erosion products built into additional ridges at the point and along the east side. Two miles west of the point the ridges show a complex arrangement over a broad area, but along a line drawn from Lydd to the point of the ness the succession of ridges is fairly regular. During the reign of Elizabeth, the distance from Lydd church to the extremity of the point was three miles, according to Redman[23]. In 1860, as shown by sheet No. 4 of the Geological Survey of Great Britain, this same distance was nearly four miles, indicating that the point advanced seaward about one mile in a little less than three centuries, which is equivalent to an annual advance of a little over 6 yards. Redman[24] studied the rate of advance as indicated by various lines of evidence accessible to him in 1852 and concluded "that the average annual increase, during two centuries has at least amounted to nearly 6 yards." Drew[25] found that from 1794 to 1860 the annual advance was about $5\frac{1}{2}$ yards.

There are about 25 beach ridges shown on Drew's map (Sheet 4, Geological Survey of Great Britain), as crossing the last mile of the distance from Lydd to the point of the ness. Although

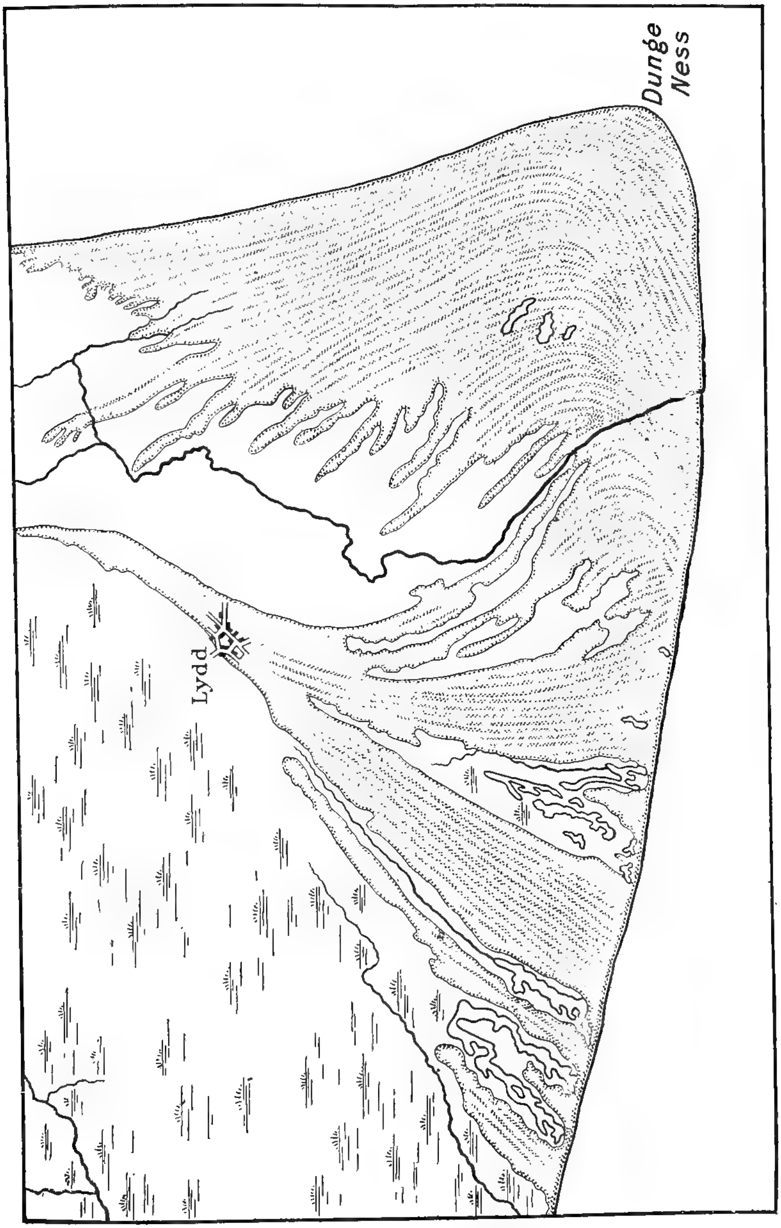


FIG. 130. — The Dungeness cuspate foreland, showing shingle beach ridges and swales.

Drew states that south and southeast of Rye the ridges are more numerous than could be shown upon the map[26], in discussing the changes near the point of the Dungeness he says that he "inserted all the 'fulls' or shingle ridges on the previously featureless Ordnance map[27]." Gulliver[28] counted twenty-three "successive shorelines" on the east side of the ness between Lydd and the sea, and as the ridges there cover a breadth of about a mile, and are shown by the Ordnance map to be between 20 and 25 in number, it would seem fair to assume that near the point of the Dungeness one ridge was built on an average every 11 or 12 years. It should be noted that some of the ridges, especially those closest to the point, are short, and that they are formed of material easily and rapidly secured from the south side of the ness which has long been suffering active erosion; both of which facts would lead us to expect an unusually rapid development of ridges near the point. That this has been the case is suggested by Redman's observation in 1852 that the point had advanced with unusual rapidity during the two years previous to his study[29], although the period is too short to be very significant. One of the coast guards stationed on the south shore of the ness informed me that the sea had removed their lookout house and cut that part of the coast back 50 feet within recent years, while the east side of the ness was advancing about 20 yards annually. This is in apparent disagreement with Gulliver's statement in 1897 that recent observation indicated an annual advance of but $1\frac{1}{2}$ yards[30]; but both figures may be correct for limited periods.

The second edition of Lewin's "Invasion of Britain by Julius Cæsar[31]" contains an interesting map, reproduced by Burrows[32] in his volume on Cinque Ports, which shows the location of marshy lands on the Dungeness reclaimed previous to the 14th century. From this map it appears that the Denge Marsh, east of Lydd, was dyked about 774 A.D. Since this marsh could hardly have come into existence until at least one beach ridge had formed to the east of the present position of the marsh to shut out the vigorous waves incident to such an exposed locality, it would appear that the ridges east of Lydd, already stated to be 23 in number, have been formed in the interval between a date previous to 774 and the present time. This would mean an average of about 50 years for the construction of each

PLATE L.

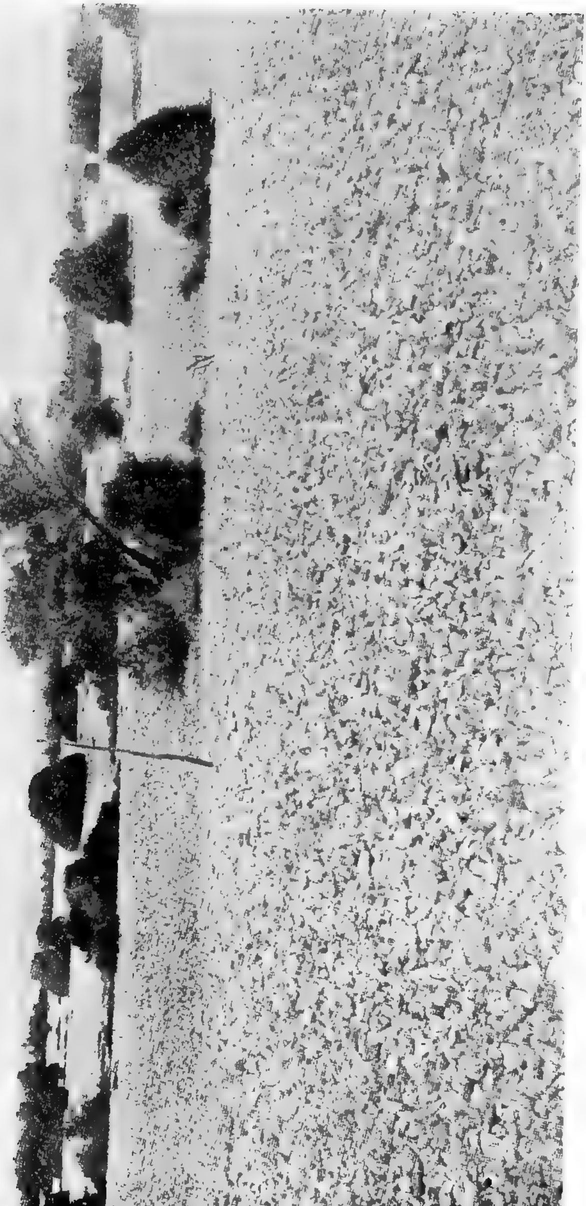


Photo by R. Hermann.

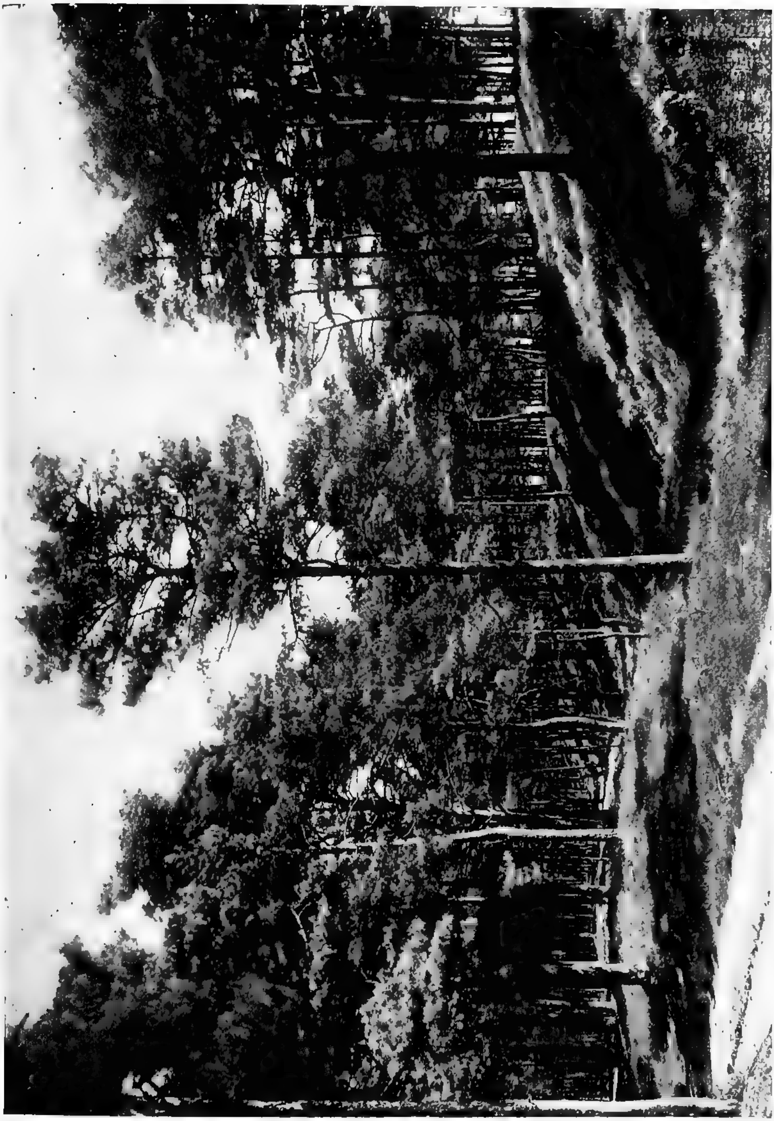
Shingle beach ridges on the island of Rügen, Germany.

beach ridge. Drew[33] considers that the region east of Lydd was open sea up to the tenth or eleventh centuries, and while his arguments are not wholly conclusive on this point, it may be noted that on the basis of his interpretation each ridge required not more than 35 to 40 years for its construction. That the older ridges southwest of Lydd are of considerable antiquity is indicated by the weathered character of their component pebbles[34].

An attempt has been made to show that the Dungeness did not exist at all in the time of Julius Cæsar, and Appach[35] gives a map of the supposed condition of this part of the English coast in the year 55 B.C. upon which the foreland does not appear. Should this contention be valid, then the ridges of the Dungeness, numbering in 1860 at least 135 according to a map which probably does not show the full number, must all have formed within an interval of little more than 1900 years; or at an average rate of one ridge in 14 years. There are ample grounds for rejecting Appach's conclusions, however. He did not properly understand the processes by which the Dungeness was formed, and his methods of reasoning are unconvincing. The fact that certain towns formerly seaports are now far inland, upon which he bases some of his arguments in favor of the recent construction of the foreland, is readily explained by Lewin's map which shows navigable bays back of the beach ridges of Dungeness point. The towns were located upon bays, which have since silted up and been converted into dyked marshes. Roman remains are found extensively over Romney Marsh which occupies the northern half of the foreland, proving that a large part of the Dungeness was completed and under cultivation in Roman times[36]. Robertson[37] has likewise demonstrated that much of the Dungeness existed at this ancient period. This means that the construction of the beach ridges of the entire foreland occupied an unknown length of time, certainly greater than 2000 years, and probably very much greater.

The available data accordingly indicates that the rate of beach ridge formation on the Dungeness foreland has varied greatly at different times, the average rate over a number of years rising as high as one ridge every 11 or 12 years at certain times and places, and dropping at least as low as one ridge in 40 or 50

PLATE LI.



Forested dune ridges on the Darss cuspate foreland, Germany, showing marked inequality in altitude of crestlines.

years elsewhere. One must fully recognize, however, that even at a given place and period the building of ridges is neither uniform in rate nor necessarily continuously forward. In a series of ridges formed at the average rate of one every 12 years a certain ridge may have required half a century or more for its completion, several other ridges may all have been built within a decade, while still others may have been built and later destroyed by a temporary erosion, thereby lowering the average rate of ridge formation for the series as a whole. For this reason, rates of beach ridge formation based on data covering very short periods are not of much value.

Taking all the facts into consideration I am inclined to believe that an average rate of one ridge constructed every 20 to 40 years is probably a reasonable figure for the Dungeness as a whole.

Darss Cuspate Foreland. — With the exception of Cape Canaveral, the finest example of a cuspate foreland composed largely of dune ridges which it has been my good fortune to see, is the Darss foreland northwest of Stralsund on the Baltic coast of Germany. Several former islands are here tied to each other and to the mainland by a complex tombolo, which has been prograded in front of the principal island (the Alt Darss) to form a triangular cuspate foreland (the Neu Darss) measuring from 7 to 10 kilometers (4 to 6 miles) on each side. Northeasterly and easterly moving beach drifting, possibly aided by other currents, transported débris which wave action built into a series of beach ridges, the axis of each ridge trending first northeast and then eastward. After each beach ridge was constructed dry sands from the shore were blown upon its crest by the winds until it rose into a dune ridge from one to several meters in height. As the foreland grew northward into the Baltic, erosion along its western side removed large portions of the ridges in that direction while redeposition of the eroded material on its northern side accelerated the northward advance. So much of the western ends of the ridges has been lost by erosion that the east trending portions alone remain to make up most of the resulting truncated cuspate foreland (Fig. 131).

Unlike the barren shingle ridges of the Dungeness, the dune ridges of the Darss are well forested (Plate LI), and the "Darsserwald" is now protected as a hunting preserve for one of

the German princes. The crests of the ridges rise 15 to 25 feet above the adjacent swales in places, and occasional ridges and a number of individual dunes reach a greater altitude. Most of the ridges do not exceed a height of 10 feet above the deepest parts of the swales, and perhaps the greater number fall short

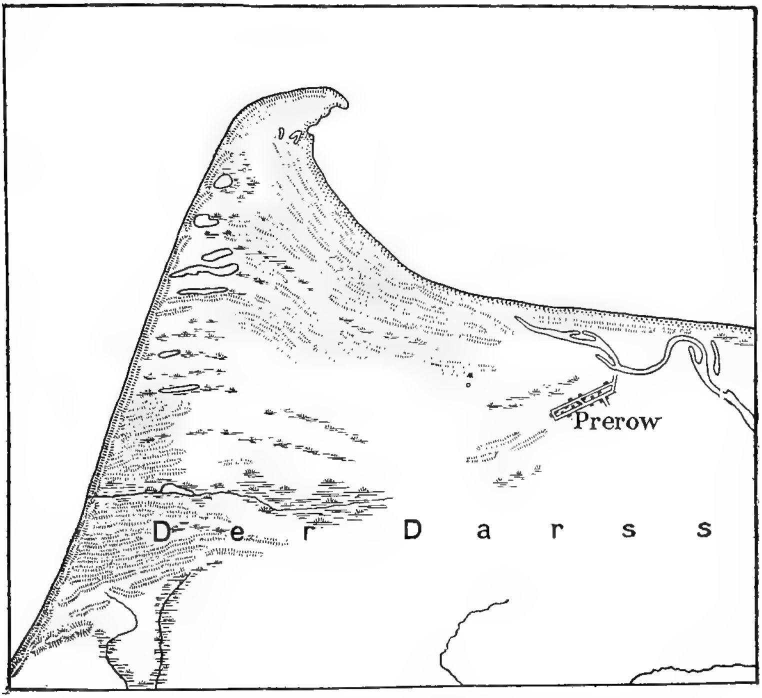


FIG. 131. — Dune ridges of the Darss cuspate foreland, Germany.

of 6 feet. Some of the swales are deep enough to contain long narrow ponds, others are marshy, while still others differ from the pine covered ridges in having fewer trees and a grassy bottom. Ordinarily the ridges are from 75 to 150 feet apart, but this distance varies greatly in different parts of the Darss, swales between 500 and 1000 feet in breadth being known. The roads through the forest are sandy, and where they are cut through the higher ridges one occasionally sees a good exposure of cross bedded dune sands, the surface layers being bleached by weathering in the ridges earliest formed; but ferns and

other vegetation usually carpet the forest floor and conceal the sand, making the region one of great beauty. There is little in the forest covering to remind one of the scrub palmetto and occasional palms of Cape Canaveral; but in spite of the contrast in vegetation, the forms of the dune ridges and swales, the variation in ridge height and spacing, and the greater weathering of the sands in the older dunes, constantly reminded me of identical features observed in the Canaveral ridges only a few months previously.

The Darss has been briefly described by Braun[38] and at great length by Otto[39]. The excellent essay of the latter author, entitled "Der Darss und Zingst: Ein Beitrag zur Entwicklungsgeschichte der Vorpommerschen Kuste," is based upon a comparative study of ancient and modern maps and detailed field investigations; and the author discusses at length the preglacial conditions of the region involved, the effects of glaciation and of post-glacial changes of level upon the coastal topography, and finally the more recent morphological changes of the coast including the development of the dune ridges. Unfortunately it is sometimes impossible to follow all of this author's arguments, because he commits the too common error of locating important features and describing essential measurements in terms of unimportant local roads, property boundaries, etc., the names of which do not appear on any maps in his report nor on any other maps available to the ordinary reader.

Otto's description of the dune ridges[40] is open to the criticism just mentioned; but I understand from the text that there are 121 dune ridges distinguishable in passing from south to north along the western side of the Darss, only a part of which number are indicated on the German topographic map of the area. Historical evidence proves that the coast has advanced 1300 feet (400 meters) in 200 years. Using this figure as a basis for calculation, and making some allowance for the fact that the younger dune ridges were probably built more rapidly than the older ones, Otto concludes that 3000 years is the shortest possible time in which the 121 ridges could have been constructed[41]. This would mean an average of at least 25 years for the construction of each ridge. Otto allows 1000 years additional for the formation and subsequent destruction of some

older ridges at the immediate base of the foreland, and thus arrives at the conclusion that the submergence which initiated the period of dune ridge formation (the " Litorinasenkung ") occurred at least 4000 years ago, or as early as 2000 B.C. If Keilhack[42] is more nearly correct in his opinion that this period of submergence occurred 7000 years ago, as seems probable to the writer for reasons which will subsequently appear, then the construction of the 121 ridges of the Darss occupied something like twice the minimum period assigned by Otto, and the average time for building each ridge would be nearly 50 years.

Swinemünde Tombolo.—The magnificent series of dune ridges, which make up the complex tombolo * connecting the islands of Usedom and Wollin some distance east of the Darss has been mentioned in many German works dealing with sand dunes, and is described at considerable length in Solger's " Dünenbuch[43]." A strait some eight miles or more in width formerly separated the two islands. Northerly winds blowing across a broad stretch of open water would drive upon the converging shores of the islands vigorous waves, which would in turn cause active beach drifting southeastward along the northeast shore of Usedom and southwestward along the northwest shore of Wollin. Two spits began to advance into the strait, the western or Swinemünde spit trending nearly due south along the east shore of Usedom, while the eastern or Misdroy spit extended itself in a more westerly direction across the strait, being strongly recurved southward at the point. The Swinemünde spit was then extensively prograded to form a beach plain by the addition of some 80 dune ridges to its seaward side, the Misdroy spit meantime advancing by gaining 150 successive recurved points at its western end while its seaward side was being retrograded. When the strait was nearly closed, erosion truncated the northern end of the Swinemünde beach plain, cut back the mainland shore of Usedom some distance, and possibly continued the previous truncation of the Misdroy recurved points. There followed a prograding of both the Swinemünde and Misdroy areas, by which

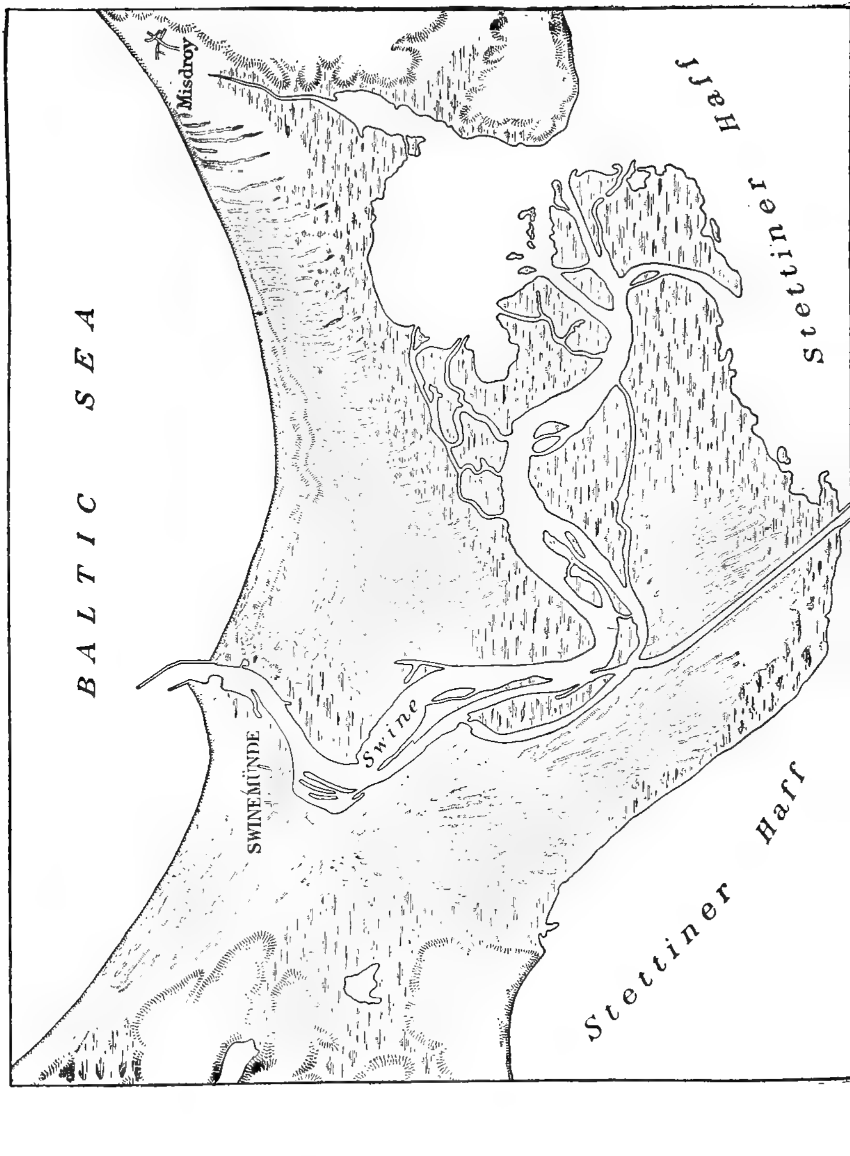
* The fact that a narrow stream passes between the islands by a channel eroded across some of the dune ridges does not alter the fact that the islands are essentially connected by a beach plain which is continuous just below water level, even if interrupted by the stream at the surface; hence I have called the combined complex spits a tombolo.

PLATE LII.

Road-cut through dune ridge at the back of the present shore at Daytona, Florida, showing bleached surface zone.

additional series of 30 and 40 dune ridges respectively were added to the northern sides of the almost united spits. Erosion slightly truncated these later ridges, and a third and last series was then added, bringing the completed tombolo to its present form. A narrow stream, the Swine, which flows alternately northward and southward is all that remains of the former strait, and it has so far shifted its position as to cut a great meander scarp into the oldest series of the Swinemünde ridges, as is clearly shown by the map (Fig. 132).

Keilhack has made a careful study of this remarkable series of dune ridges, and has published his results in a valuable essay on "Der Verlandung der Swinepforte." He found the distance between ridge crests to vary from 130 to 150 feet where they were closely spaced, and from 330 to 460 feet where they were farther apart. In altitude the older ridges usually do not exceed 25 feet absolute elevation, but the earliest ridge formed in the third or last series reaches a height of 65 feet or more[44]. Of especial interest are Keilhack's observations on the comparative weathering effects in the three systems of dune ridges[45]. The dunes of the latest series are practically unweathered and retain the normal light color of the beach sands from which they were formed; they are, therefore, called "white dunes." Dunes of the next older series show a thin surface layer of bleached sand, below which the sand is colored yellow by limonite; these are known as the "yellow dunes." Finally, the oldest dunes have a thin surface layer of humus from less than an inch to an inch or more in thickness, below which is the bleached sand zone from 1 to 1½ feet thick. Beneath the bleached zone the sand grains have a coating of brown limonite, and may even be locally cemented by this material into a soft ferruginous sandstone. These "brown dunes" must have existed essentially as we find them for a long period of time in order to experience such pronounced weathering effects. The formation of the bleached zone is attributed to the leaching action of atmospheric waters carrying CO_2 and humus acids, by means of which all iron is removed from the upper foot or eighteen inches of each dune ridge. How clearly the white surface band is contrasted with the darker sand below when exposed in cross section, may be seen in Plate LII, which represents a road-cut through an old dune ridge back of the present shore at Daytona,


BALTIC SEA
Misdroy
Stettiner Haff
Swine
SWINEMÜNDE
Stettiner Haff

Florida. The weathering phenomena characteristic of the yellow dunes of Keilhack was clearly evident in the older dunes of the Darss, but I saw no such advanced stages of alteration as that author describes for his brown dunes. I am, therefore, inclined to agree with Otto[46] that the oldest preserved dune ridges of the Darss are not so ancient as the oldest ridges near Swinemünde. Perhaps the ridges earliest formed near the base of the Darss and later eroded[47] were more closely similar to the brown dune ridges of Keilhack.

Through a comparison of reliable maps Keilhack has been able to show that between the year 1694 and the beginning of the twentieth century the shore west of the northern outlet of the Swine was prograded nearly one mile (1500 meters), while elsewhere the advance was less marked. Since 1694 six dune ridges have been formed, or an average of one ridge in every 35 years. The author then points out that this figure agrees so remarkably with the figure found by Brückner for a periodic climatic oscillation, that one cannot well refuse to accept Solger's opinion in favor of a genetic connection between the formation of parallel dune ridges and this climatic period. He, therefore, accepts 35 years as the time represented by each ridge, and derives a chronology for the entire tombolo. East of the Swine, the 150 ridges of brown dune forming the original Misdroy spit would require 5200 years; the 40 ridges of yellow dunes which followed would demand 1400 years; and the 7 or 8 ridges of white dunes about 300 years additional; making a total of 7000 years for the entire series of dune ridges on the Misdroy side of the tombolo. The number of ridges on the Swinemünde side is much less, but the record there is assumed to be less complete.

To the 7000 years derived in the manner above indicated, Keilhack would add an unknown number of years representing two erosion periods which separated the three systems of dune ridges. The evidence for two erosion periods, distinct from the periods of prograding, is not convincing, and Keilhack's discussion of this question does not appear to be consistent. To account for the truncation of the northern ends of the Swinemünde brown dunes previous to the formation of the next following series of yellow dunes, he invokes a subsidence of the entire district, amounting possibly to as much as 6 to 10

feet, which would decrease the supply of marine sand for dune building and favor erosion[48]; at the end of the erosion period, estimated as 1000 to 2000 years in length, it seems that re-elevation is considered a probable though not necessary cause of the resumption of ridge building which resulted in the next series of yellow dunes[49]. Inasmuch as coastal subsidence is appealed to in order to account for the cessation of ridge building and the initiation of erosion on the Swinemünde portion of the area, it would seem natural to expect that the subsidence would effect the same changes on the Misdroy hook just across the Swine. But since the Swinemünde hook has but 80 ridges, and the Misdroy hook 150 ridges in the oldest series, it is evident that according to Keilhack's interpretation the Misdroy hook must have continued to advance for 2400 years after subsidence is supposed to have arrested the advance of the Swinemünde hook; indeed, Keilhack specifically states that the erosion which truncated the Swinemünde hook may very well have occurred during the same 2400 years that the Misdroy hook was still advancing[50], apparently not realizing that this invalidates his previous arguments in favor of repeated depressions and re-elevations of the area as a cause of alternate periods of shoreline erosion and deposition. To account for the cessation of the building of the yellow dunes, their truncation by erosion, and the later building of the white dunes, Keilhack imagines a second movement of subsidence, introducing an erosion period some hundreds of years long, followed probably by a slight elevation which occurred between 1500 and 1600 A.D. and exposed great masses of sand on a wide beach which the wind could build into the especially high dune ridge which marks the beginning of the white dune system[51]; but on the following page of his essay he states that the eastern half of the Swinemünde-Misdroy region continues to be eroded up to the present day. Thus this author invokes coastal subsidence in order to account for shoreline erosion, yet recognizes such erosion following coastal elevation.

The reasons for rejecting the oft-repeated opinion that shoreline erosion implies coastal subsidence have already been discussed at some length. In the opinion of the present writer all of the phenomena described by Keilhack as characteristic of the Swinepforte dune ridges are readily to be explained without invoking

any changes in relative level of land and sea. Beach ridges and dune ridges have in the past been built forward at one place and truncated in another simultaneously, just as the Dungeness is today having its ridges of shingle cut away on the south side and built forward on the east; or as Cape Canaveral is being eroded on the east, and prograded on the south; or, indeed, as the white dunes near Swinemünde have been built forward in the same time that the closely adjacent coast was cut back. In fact, the erosion at one place causes, or at least accelerates, the forward building at another by increasing the supply of shore débris. It is to be expected that progressive addition of beach or dune ridges will in time so change the outline of the shore and hence the intensity and direction of marine forces, that the profile of equilibrium on adjacent parts of the shore will be disturbed, and erosion will replace deposition at certain points, without any change in land or sea level and without any profound revolution in the nature of the marine forces operating on the shore. The equilibrium of a shore profile is a very delicate thing, and it may very easily be so disturbed that an excess of erosion replaces a former excess of deposition.

It is highly probable that much if not all of the erosion of dune ridges which occurred in the Swinepforte district took place while dune ridges were forming in other parts of the area; and that, therefore, no additional time is to be allowed for these erosion intervals. Keilhack recognized the uncertainty of the erosion intervals, and, therefore, permitted his estimate of 7000 years to remain unchanged, merely stating that the time interval since the Litorina submergence which introduced the period of ridge building must be more than 7000 years.

We may accept Keilhack's estimate of 35 years as the average time required for the construction of each of the six ridges of white dunes formed since 1694, without agreeing to the correlation of dune ridge development with Brückner's climatic cycle, or to the proposed chronology of the older dune ridges. We have already seen that the physical forces which control the growth of successive beach and dune ridges are so important in magnitude and so variable in their activities that they would scarcely be materially affected by the very moderate climatic changes of the 35 year period. It is true that Krüger[52] in his study of "Sturmfluten an den deutschen Küsten der

westlichen Ostsee" reaches the conclusion that periods of frequent "storm tides" alternate with periods in which their occurrence is rare, and that these periods correspond in a general way with the dry and wet periods respectively of the Brückner cycle. There are, however, striking exceptions to Krüger's rule which cast some doubt on its value and certainly invalidate it for use in establishing a beach ridge chronology. There seems to be no escape from the conclusion that the supply of sand, the intensity and frequency of great storms, the length and position of the ridges, and other controlling factors have varied so greatly during the building of the Swinemünde-Misdroy tombolo that the average time for ridge building has been very different at different places and at different periods. Thus it is not impossible that the 150 short recurved points of the original Misdroy spit were built in nearly the same length of time as the 80 longer ridges which were added to the front side of the Swinemünde spit, even though the cutting of the meander scarp in the Swinemünde series suggests that the Misdroy spit may have added a few of its recurved points after the Swinemünde ridges were completed, thereby deflecting the Swine against the latter.

The Swine has built a beautiful delta into the Haff south of the tombolo, and it seems probable that it carries an appreciable amount of débris into the bay on the north when it flows in that direction. If so, wave action should utilize this débris to prograde the shore with unusual rapidity near the Swine mouth. The existence of moles or jetties on either side of the mouth may also tend to check longshore transportation and to accelerate prograding in that vicinity. As shown by Figure 132, there is a delta-like projection of the dune ridge series at the mouth of the Swine, where formerly an embayment existed as shown by the older ridges; and Keilhack states that the moles at the mouth of the Swine have made the shore build forward there much more rapidly than usual during the last two centuries[53]. The six dune ridges built within this same period, and used by Keilhack as a basis for his calculations, may, therefore, represent a much smaller time interval than six ridges of the older series. Many of the latter may have required an average of 50 years or more for the construction of each ridge.

Enough evidence has been presented to show the impossi-

bility of building up any accurate chronology on the basis of beach ridges or dune ridges. On the other hand, it appears that a large series of extensive ridges must represent a long time interval, and that 25 to 50 years is not an improbable figure for the time required to build such prominent ridges as are characteristic of the Dungeness, Darss, and Swinepforte areas. Beach and dune ridges, therefore, have a great value in acquainting us with the order of magnitude of the minimum time involved in their construction, even though they cannot furnish more precise data.

Beach Ridges as Records of Changes of Level. — A well-developed series of beach ridges may have a high value as evidence of former changes in relative level of land and sea, or of coastal stability. If there is a gradual emergence of the land during the development of the ridges, it would seem that the crests of older members of the series should be found at progressively higher elevations above water level; whereas continued submergence should be indicated by a decrease in crest altitude as one passes inland from the modern ridges. Coastal stability, on the other hand, should be recorded by a general agreement of ridge crest altitude throughout the series.

If applied with discrimination and with a full understanding of the different conditions which determine the altitudes of beach ridges, the above principle may throw valuable light on the interesting questions relating to past changes in the level of land and sea. Its indiscriminate and uncritical use will often lead to erroneous conclusions. It behooves us, therefore, to take cognizance of certain fundamental facts concerning the formation of beach ridges, and to note in what ways they may affect our judgment in interpreting the significance of crest altitudes. Again it will be most convenient to state the facts categorically, and comment on them as may seem desirable.

1. The terminal points of recurved spits normally descend toward their distal ends and pass under the water level, as has previously been shown. In a compound recurved spit, therefore, it will often be found that the ridges back of the present shore, representing successive recurved points, are materially lower than the modern beach ridge. It is difficult to see how any one could regard such difference in ridge crest elevation as an evidence of coastal subsidence; yet it has been so regarded

by several observers. One should clearly realize, however, that the point of a spit which curves back into more protected and quieter waters thereby escapes that direct impact of the larger waves which is necessary to heap up the sand or gravel to the greatest altitudes; and that the failure of an adequate supply of débris near the terminus necessitates a low embankment in any case. Observation will suffice to show that where the points of such spits are lengthening from year to year, they are built low in the first place, and do not acquire their low level by subsidence.

2. Beach and dune ridges of great linear extent normally vary in altitude along their crests. If they possess free ends, they usually descend more or less gradually and pass under the water; for while they may not recurve into quieter water, such unattached ridge ends resemble spits to the extent that the supply of débris at their distal points is insufficient to build up a submarine embankment and raise it to a considerable elevation above sealevel. It is not necessary, for example, to regard the descending southern end of the oldest Swinemünde ridges[54] as an evidence of coastal submergence. Variations in supply of material, in exposure to wave action, in depth of offshore bottom, and in other factors may cause a marked variation in crest altitude anywhere along the course of the dunes. On the Darss foreland, where observations indicate long-continued coastal stability, a large number of the older east-west dune ridges are low in the central part and high at either end.

3. Successive beach and dune ridges normally differ from each other in altitude of crest line. This follows from what has already been said regarding the origin of such ridges. A temporary excess of shore débris may cause a new ridge to form before the earlier one behind it had acquired any considerable altitude; and the new ridge may rise to a great height before the development of a still later ridge checks its growth. Temporary retrograding of the shoreline may combine several low ridges into one high one, while earlier and later ridges remain of moderate altitude. The great variability of the marine forces causes the successive positions of the shoreline to be maintained for unequal lengths of time, and to have unequal quantities of shore débris cast into shore ridges of unequal height. It may happen that one ridge is not raised above

PLATE LIII.

Successive curving beach ridges and swales forming lines of growth on Cape Canaveral. View from Cape Canaveral lighthouse tower.

water level before another is built in front of it, but as a rule the differences in height are all to be measured above the level of the sea. Beach ridges are formed directly by the waves, and cannot, of course, exceed the height to which waves in a given exposure are capable of raising the material of which the ridges are composed. This may be only three or four feet in a sheltered locality, but very commonly amounts to 10 or 15 feet for ordinary shingle beach ridges on an open coast, and a single semi-permanent ridge like the Chesil Bank on the exposed south coast of England may reach a height of 40 to 50 feet above high water[55]. It is no uncommon thing to find beach ridges 10 feet or more in height irregularly interspersed with others less than half as high; and theoretically the difference may be as great as, or greater than, the maximum height of the ridges above water level. Practically, however, the irregular variations in crest altitude are commonly not much greater than half the altitude of the higher ridges, and in many cases are appreciably less. Goldthwait[56] has described a case in which a series of sixteen consecutive ridges having an average crestline altitude of 3.82 feet above high water contained no ridge higher than 4.64 feet nor any lower than 3.04 feet; an extreme difference of but $1\frac{1}{2}$ feet.

Dune ridges owe their height to the action of the wind, and may, therefore, rise well above the upper reach of storm waves for a given exposure. Dune ridges 65 feet high are known[57] on the shores of the Baltic, and the remarkable height of nearly 300 feet is reported from the somewhat irregular dune ridges along the coast of the Landes in France[58]. Usually, however, 15 to 25 feet is the upper limit for individual members of an extensive series of dune ridges. As is to be expected, variations in altitude among different ridges of a given dune system are greater than in the case of beach ridges, because to the variable factors affecting the initial beach ridge upon which the dunes stand are added the variable factors which affect dune accumulation, including strength and direction of the winds which move the dry sands of the upper part of the beach, and the character of the dune vegetation. Twelve dune ridges, associated with the very accordant beach ridges described by Goldthwait and mentioned above, were found by that author to vary from 3.91 to 9.04 feet in height above high water[59]. Dune ridges on the

Darss vary in altitude from 2 feet or less to 25 feet or more, measured from the bottom of adjacent swales. Among the Swinemünde dune ridges are many 3 to 6 feet high, others 25 feet or more, and one or two as high as 65 feet above sealevel. The Cape Canaveral dune ridges vary from 2 to 12 feet above high tide level. Perhaps the lowest dune ridges are merely beach ridges with the surface sands slightly disturbed by the wind.

It is manifestly impossible to regard such variations in the level of individual beach ridge and dune ridge crests as indications of elevations and subsidences of the land. Probably no one would be so bold as to imagine such a rapid and oft-repeated alternate rising and falling of the coast as would be called for by the great series of ridges of the Dungeness, Darss, Swinemünde, and Canaveral beach plains, were variations in ridge height to be regarded as proving variations in sealevel. It follows that one should be equally cautious in accepting the inequality in height of two or three ridges as a proof of changes of level; for if many ridges may acquire unequal altitudes without the aid of vertical movements of land or sea, certainly a few may do so. If we are satisfied as to the validity of this conclusion, we shall have no difficulty in realizing the fallacy of one of the lines of argument not infrequently advanced in support of theories of coastal subsidence and elevation.

4. In a given series of beach or dune ridges there is a tendency for those first formed to have a lower altitude than later members of the series. Cornish[60] was of the opinion that "it is unnecessary to invoke upheaval or subsidence to account for such difference of level," and explained the greater height of the later ridges on the ground that as a foreland builds farther and farther out into the water it offers increased obstruction to the coastal currents, thus causing them to bank up the water to a greater height and raising the level of ridge construction. We may agree with Cornish's general conclusion, yet doubt whether the level of the water is ever sufficiently affected by foreland growth to account for the phenomena in question. It is evident, however, that on a sloping bottom only small waves can operate near the shore, since waves break when entering water of a depth about equal to their height. A beach ridge built near the shore will tend to have a low altitude, for small

waves cannot cast débris to a great elevation. Waves of greater height, breaking farther seaward, finally build a new ridge in front of the one first formed, and are able to build the crest of this new member of the series to a somewhat greater altitude than that of its predecessor. Still larger storm waves may build a third ridge of still greater height; and in this manner there is produced, as the result of normal wave action on a stable coast, a series of beach ridges of increasing altitude going seaward. There can be no doubt that this history of beach ridge development has been repeated in many places along our coasts, and it is, therefore, manifestly impossible to regard a landward decrease in beach crest altitude, especially in a series of a few ridges only, as a proof of coastal subsidence.

During the early stages of beach ridge formation on a shelving sea-bottom, it is probable that the zone of ridge building is shifted seaward with constantly diminishing rapidity. The first ridge is quickly built by the smaller waves. Soon larger waves begin the construction of a new ridge in front of the first. The shoreline remains for a longer time in this new position, because no change will occur until the waves have built up from a deeper sea-bottom a ridge of sufficient height to transfer the shore activities permanently to a third position still farther seaward. We have already seen that the longer a shoreline remains in a given position, the greater is the probability that the shore ridge will be raised to a high elevation. On this account we may reasonably suppose that progressively slower advance of a shoreline often helps to produce a series of beach ridges whose crest altitudes decrease in a landward direction.

A further cause of normal decrease in altitude of progressively older beach ridges is probably to be found in the greater weathering to which the older members of the series have been subjected. In the course of many centuries it seems certain that a ridge of gravel loosely piled up by the waves must become somewhat compacted; while sand ridges will be very slowly worn lower under the constant attack of rains and other agencies of weathering. It can hardly be supposed that in the course of a few thousand years such changes in crest altitude would be very pronounced; but we may fairly assume that they would be appreciable, and might therefore serve to augment similar differences due to other causes.

Dune ridges formed on low beach ridges, and dune ridges that have had only a short time in which to accumulate by reason of a rapid prograding of the shore, or that have been acted upon by the weather for thousands of years, tend to have a low crest altitude. From what has been said regarding beach ridges it follows, therefore, that successive dune ridges with diminishing crest heights going inland may be a normal feature of a stable coast, and that they are no more to be regarded as proofs of coastal subsidence than are beach ridges showing similar relations of crest lines.

It should be fully understood that beach and dune ridges of progressively decreasing altitude landward are normal, but by no means necessary, features of a prograding shore. They are more apt to characterize the earliest stages of shore prograding, and we must be prepared to find the oldest members of a large series of ridges, or all the members of a very small series, showing the phenomenon in question at various places along a shore which has experienced no change of level since ridge building began. On the other hand, the conditions which determine shoreline development are so complex and are subject to such variations that one cannot expect to find a simple, regular decrease in crest altitude as a common feature of all beach and dune ridge series. On the contrary, it is only under favorable conditions that the tendency to produce such regular differences in altitude is not masked or completely overcome by other forces. We shall find instances in which rapid prograding of the shore has produced a series of low ridges next the present shoreline, while older ridges have a higher average elevation.

5. Beach ridges are more valuable than dune ridges in determining changes of level or coastal stability. This follows from the fact that dune ridges show greater local variations in height than do beach ridges, as is explained in paragraph No. 3 above. It is clear that safe conclusions as to past moderate variations of relative sealevel or past coastal stability cannot be so readily based upon ridges which may show wide differences of level due to causes independent of vertical changes in the position of land or sea, as they can upon ridges which normally vary within much narrower limits.

6. Beach ridges and dune ridges must be regarded as incomparable features, when one is seeking to determine the possi-

bility of past changes of level. It is not permissible to compare an older series of beach ridges with a later series of dune ridges, or *vice versa*, and to infer subsidence or elevation if the average heights of the two dissimilar series are unlike. The force which predominates in dune building is not the same as the force which predominates in beach building, and there is no reason why the two forces should build ridges of similar height. On the contrary, dune ridges are built upon pre-existing beach ridges, and must, therefore, exceed the latter in altitude. Near Sandhammaren on the southern coast of Sweden a magnificent series of beach ridges is bordered seaward by a higher series of dune ridges; but the theory that this part of the Swedish coast is subsiding is disproved by evidence which I will present in another connection. A similar relation of beach and dune ridges on the coast of eastern Canada has been cited by Ganong as an evidence of coastal subsidence, although Goldthwait[61] has shown that the later and higher dune ridges of this region rest upon gravel beach ridges of the same height as the older beach ridges. Beach ridges may properly be compared as to altitude with other

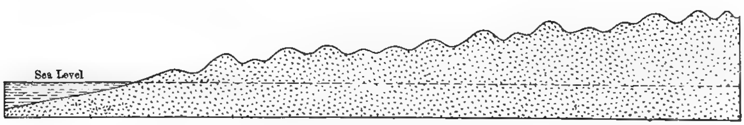


FIG. 133. — Beach ridges indicating coastal emergence.

beach ridges underlying dune ridges, when their surfaces are sufficiently exposed for this purpose; but never with the superimposed dune ridges themselves.

7. Both beach ridges and dune ridges have a distinct value as records of changes of level or of coastal stability, notwithstanding the restrictions mentioned above. A large series of beach ridges which may show irregular variations in heights of individual crests but which is characterized in addition by a gradual landward increase in the average height of the ridges, or in the heights of the principal ridges (Fig. 133), is strongly suggestive of emergence. If the seaward members of the series have a considerable height, indicating that they have more or less nearly attained the maximum elevation which waves in that exposure can give to ridges, while the older ridges have a

much greater height, the evidence may be said to furnish satisfactory proof of elevation. I have found such ridges on the northern shores of the Baltic Sea, where independent evidence indicates progressive emergence of the land. Care should be taken, however, in employing this line of evidence in those cases where the prograding of the shore materially reduces the width of the water body upon which the ridge-making waves are developed; for the size of the waves will thereby be reduced, and the seaward members of the ridge series will decrease in altitude independently of coastal emergence.

We have already seen that a few beach ridges exhibiting a landward decrease in crest altitude is a normal feature of a stable shore, and therefore must not be regarded as an evidence of coastal submergence. It is even probable that a large series of such ridges may be characterized by the same landward decrease of average crest altitude, due to a gradual seaward increase in depth of water and size of waves, and to other factors favoring greater crest height in the later ridges. If an extensive series of beach ridges descending landward could be traced to a considerable depth beneath the surface of a salt-marsh peat deposit composed of high tide vegetation only, which had protected the ridges from destruction by extending over them as they sank lower and lower (Fig. 134), coastal submerg-

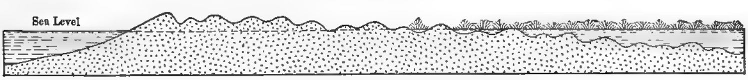
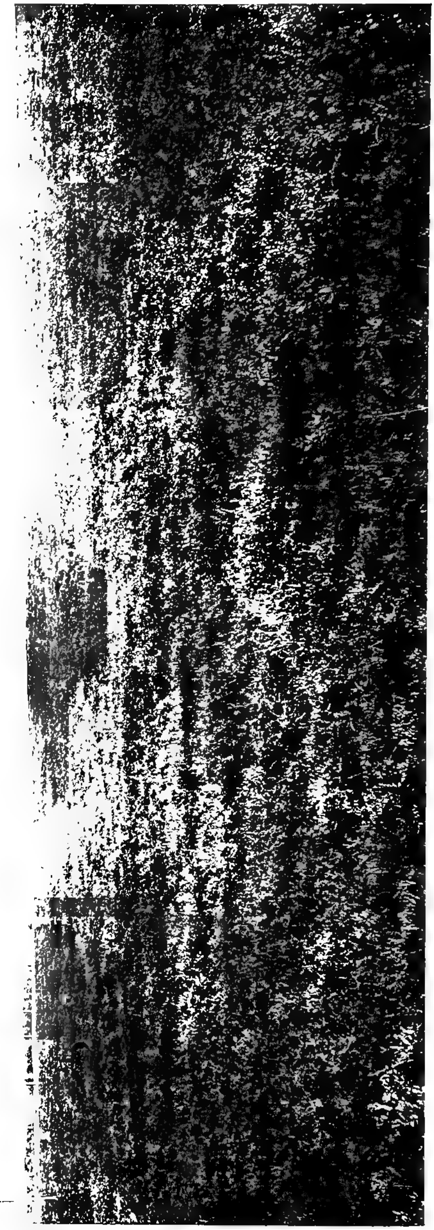


Fig. 134. — Beach ridges indicating coastal submergence.

ence could be inferred with reasonable certainty. A few ridges at a shallow depth in the marsh would not be satisfactory evidence; for normal wave action on a stable shore might fail to raise the initial ridges above sealevel, while marsh deposits might later protect them from destruction by the lagoon waves. It is very seldom that the conditions which render it safe to employ beach ridges as an evidence of coastal submergence exist. In all of the cases which have come to my attention where a landward decrease in ridge crest height has been used as a proof of submergence, such use has not seemed to me justifiable, for the reason that the phenomena described might equally well be explained as the normal product of wave action

PLATE LIV.



Level surface of Skanör peninsula, southwestern Sweden. A beach plain consisting of closely spaced beach ridges.

on a stable coast. It may reasonably be doubted whether beach ridge development often takes place on a subsiding coast, since subsidence favors marine erosion, and is highly unfavorable to the prograding of shorelines.

Where a large series of beach ridges show throughout about the same average crest altitude, or about the same altitude for the principal ridges, coastal stability is strongly indicated. If the older and later ridges are both about as high as the present waves could be expected to build them, the evidence in favor of long continued stability may be regarded as conclusive. There are two hypothetical cases which might lead to an erroneous conclusion, but it is probable that danger of error from this source would be eliminated by careful observation. One may imagine that on a rising coast where the earliest ridges are of small altitude and the later ridges progressively higher, the amount of elevation might just be sufficient to raise the crests of the first ridges into the same horizontal plane with the crests

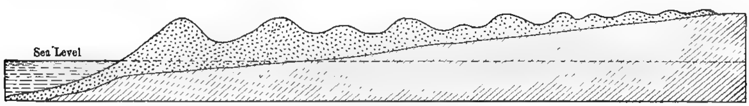


FIG. 135. — Hypothetical case in which beach ridges on a rising coast may give a false indication of stability.

of those formed later (Fig. 135); and a careless observer might argue in favor of coastal stability because of the resulting equality of crest heights. But since we are not apt to find high beach ridges with very narrow bases, while the low ridges formed in shallow water are characteristically narrow, comparison of the older and later ridges formed in the manner indicated should reveal the fact that those first formed are really low ridges raised high above the plane in which they must originally have been constructed. This is made clear by Figure 135. The case is improbable, not merely because the rate of emergence must be just enough to give the required equality of crest altitudes, but also because a progressively emerging shore favors the repeated development of small ridges rather than ridges of constantly increasing height.

A second case may be imagined in which progressive submergence carries the crests of older, high ridges nearer to water

level, thereby bringing them into the same horizontal plane as the crests of successively lower ridges formed later. Thus, as shown in Figure 136, one might infer coastal stability from equality of ridge crest altitude in a region which had really experienced progressive submergence. The true history might be suspected from the fact that an increasing proportion of the older larger ridges was below marsh level or the level of lagoons caused by

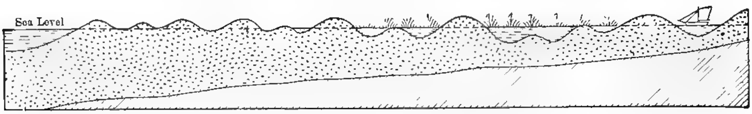


Fig. 136. — Hypothetical case in which beach ridges on a sinking coast give a false indication of stability.

the submergence. This hypothetical case, is, however, even more improbable than the one supposed above, since it involves not only a special rate of subsidence and the building of the largest ridges in the shallowest water where only small ridges are to be expected, but also because submergence tends to prevent ridge building entirely and to favor the erosion of the coast. As shown by the figure, the formation of the smaller ridges demands an increasingly extensive aggrading of the deeper off-shore bottom, a process to which submergence is distinctly unfavorable.

Widely spaced older beach ridges rising above marsh level back of a later series, thereby giving a superficial appearance of the conditions represented in Figure 136 must not be regarded as an indication of subsidence, since such ridges may have been formed with wide spaces of water between them in the first instance, and the lagoons converted into marshes at a later date. Several ideal profiles through beach and dune ridge series formed on stable coasts are shown in Figure 137.

Two ridges (Fig. 137 *c*) of similar altitude may be sufficient to prove long continued coastal stability, providing they are so high as to preclude the possibility that the earliest one was built much higher and later carried down by subsidence, and providing also the older one is manifestly not a small initial ridge raised to its present height by coastal elevation. In addition, there must be some means of proving the lapse of a long interval of time between the building of the two ridges. A case of this kind is

presented by Nantasket Beach, Massachusetts, and has been fully described by Johnson and Reed.[62]

An extensive series of dune ridges may furnish reliable evidence of essential coastal stability, if their formation has evidently required so long a period of time that any marked change

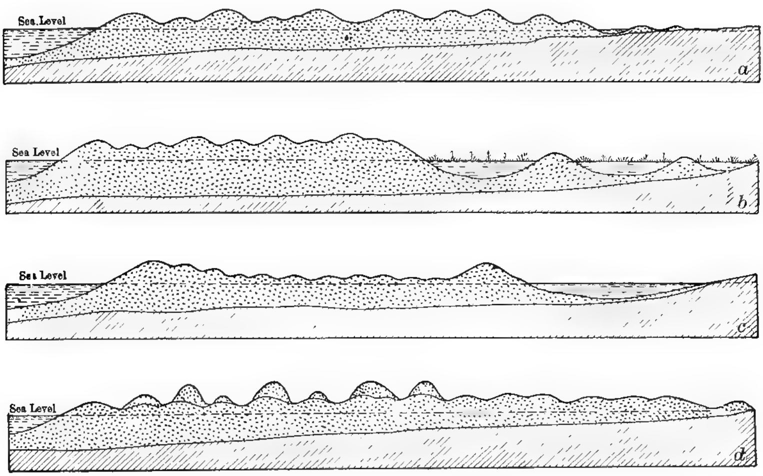


FIG. 137. — Types of beach ridges formed on a stable coast.

(*a*) Earliest beach ridges lower because of shallow water nearest the original shoreline.
(*b*) Similar to *a*, but older ridges isolated in marsh.
(*c*) Central ridges low because of rapid prograding to present zone of wave action, where the tendency to prograde is much less pronounced.
(*d*) Later ridges with greater average height than older, because former are dune ridges surmounting beach ridges, while latter are unmodified beach ridges.

of level must of necessity have resulted in a pronounced difference in crest heights recognizable in spite of individual variations in ridge altitude. For example, if the members of an extensive system of dune ridges vary in original height from 3 to 25 feet, with the exception of occasional abnormal individuals which are manifestly the product of special conditions and which may therefore be ignored; and if the average height of the older and later ridges is similar, and the building of the entire series required 5000 years; then one may safely reject a theory which would demand, for example, a continuous progressive

subsidence averaging 6 inches or a foot per century. For a subsidence at the smaller rate for the period mentioned would carry the highest of the older ridges down to sealevel and would deeply submerge the smaller ones. The fact that there has been no material change in the relation of dune crests to sea-level between the earlier and later portions of the series is sufficient indication that there has been no marked change in the relative level of land and sea. To admit the possibility of progressive subsidence of the land, we would have to assume that prograding took place in spite of subsidence, that the earliest formed ridges were built 25 feet higher, on an average, than

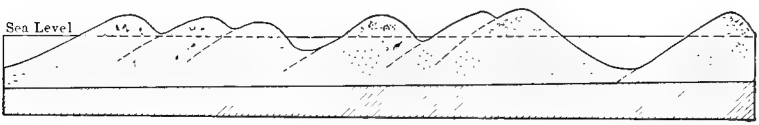


FIG. 138. — Beach ridges of equal height separated by swales of different depths due to variations in spacing of ridges.

the modern ones, and that this excess of height decreased with some degree of regularity and at about the same rate as subsidence carried the land downward; a series of assumptions difficult to grant.

8. The levels of swale bottoms, whether between beach ridges or dune ridges, is of comparatively little significance. This follows from the fact that the depth of the swales depends in large measure upon the closeness of the spacing of the ridges, which is in turn dependent upon factors not usually related to changes of level. Figure 138 will serve to make clear the fact that a series of similar ridges of equal height, built on a stable shore by a prograding process which varied in rate with variations in supply of débris by longshore currents, may be separated by swales of very unequal depth.

RÉSUMÉ

We have inquired into the origin of beach ridges and dune ridges and have found that while they are produced by waves operating under a variety of circumstances, they are not to be correlated with individual great storms. Among the types of

current action responsible for the supply of débris built into parallel ridges, longshore beach drifting resulting from waves breaking obliquely on the shore, although too commonly neglected, is believed to be one of the most important. The conditions which control the heights of beach and dune ridges have been discussed at length, as have also the conditions affecting the rate of ridge development. For our guidance in attempting to estimate the approximate time represented by any given series of beach ridges or dune ridges, certain general principles have been laid down; and an examination of the known or estimated rates of ridge formation on certain important beach plains has provided data which will be of some service in making such attempts. Finally, it has been shown that, when interpreted with caution, beach and dune ridges may furnish valuable evidence as to past changes in the relative level of land and sea; and a series of eight fundamental principles, the recognition of which is essential to a proper interpretation of such evidence, has been presented and discussed.

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CHAPTER X

MINOR SHORE FORMS

Advance Summary. — There remain for consideration a number of shore forms which are not of primary significance in a discussion of shoreline development, but which are nevertheless of much importance to the geographer and geologist, and in some cases also to the engineer. It is proposed to give some account of these features in the present chapter. Beach cusps are first discussed at much length, after which the low and ball, especially characteristic of sandy shores, are described. Ripple marks receive an extended treatment, following which rill marks, swash marks, backwash marks, sand domes, and shore dunes each in turn are briefly considered.

Beach Cusps. — Among the minor forms of the shore zone none has proved more puzzling than the cuspate deposits of beach material built by wave action along the foreshore. Sand, gravel, or coarse cobblestones are heaped together in rather uniformly spaced ridges which trend at right angles to the sea margin, tapering out to a point near the water's edge. These "beach cusps" have attracted the attention of many students, and it will be profitable for us to consider first the opinions of other writers concerning them; then of examine more carefully into their essential characteristics; and finally to criticize the various theories which have been proposed to account for their origin and development.

Previous Studies of Beach Cusps. — The earliest account of beach cusps which has come to my attention occurs in a paper on shingle beaches published by Palmer[1] in 1834. Palmer's description of the forms is very vague, but he recognized the important fact, not appreciated by all later students, that the cusps are produced by waves "driven directly upon the beach," whereas they are destroyed when "an oblique direction is given to the motion of the waves." In an unpublished thesis, "The Geology of Nahant" written by Lane about 1887, the cusps

PLATE LV.

Photo by W. S. Cooper.

Beach cusps of gravel on the shore of Carmel Bay, California.

on Lynn Beach, Massachusetts, are briefly described and their origin discussed. Lane concluded that cusps are formed by the action of waves parallel to the coast; that they have their beginnings in accidental irregularities on the beach; that they become evenly spaced as the result of some process of adjustment not clearly understood, and that the distance between cusps is in some manner related to the height of the waves and the breadth of the beach. A short abstract of this thesis was published in 1888, but contains only a brief reference to the cusps[2].

A few years later Shaler, in his popular treatise, "Sea and Land[3]," gave a clear description of the curious "ridges and furrows" occurring on shores, recognized their temporary character and the ease with which they are obliterated by wave action, and expressed the opinion that "the origin of these peculiar structures is not easily accounted for." Shaler published a somewhat fuller account of beach cusps in his paper on "Beaches and Tidal Marshes of the Atlantic Coast." A theory of origin was there proposed in the following words:

"It seems to the writer that these scallops were formed about as follows: In a time of storm the inner edge of the swash line formed by the body of water which sweeps up and down the beach has a very indented front, due to the fact that it is shaped by a criss-cross action of many waves. As these tongues run up the beach and strike the pebbles, they push them back so as to make a slight indentation where each tongue strikes. As the water goes back, it pulls out the fine material, but does not withdraw the pebbles. The next stroke of the splashing water then finds a small bay, the converging horns of which slightly heap up the fluid, making the stroke a little harder in the center of the tongue and excavating the bottom of the bay still farther. As the re-entrant grows larger and the tide rises higher, the water, as it runs up, forms a small wave, which breaks on the shore of the recess and casts the pebbles more into the form of a ridge. This action, continuing for some hours before the tide turns, serves to shape the embayment.

"It should be carefully noted that, when the swaying waters rush up into the shore scallops, the converging walls of these indentations deepen the current and add to the efficiency of its movements — a process which is essentially like that which is brought about when an ordinary wave enters into a recess of the

cliff, or the tidal undulation is crowded into an indentation such as the Bay of Fundy[4]."

In his paper on " Sea-beaches and Sand-banks " published in 1898, Cornish briefly refers to the " succession of ridge and furrow at right angles to the sea-front," and attributes the phenomenon to the erosive action of waves which are increasing in size and attempting to reduce the beach slope to a gentler gradient. A variation of the same feature is described by Cornish under the name " Shingle Barchanes." He was of the opinion that the shingle barchanes were analogous to that form of sand dune called a barchane, and considered any discussion of their origin superfluous[5].

One year later Jefferson published a paper in which he described some of the characteristic features of beach cusps and offered an explanation of their origin. Jefferson's studies were " made at a single beach (Lynn Beach, Massachusetts), though confirmed by some observations from Gay Head and Narragansett Bay." He concluded that the cusps were caused by the escape of water from behind a barrier of seaweed located near the upper zone of the beach. Occasional waves of more than average size overtop the seaweed barrier and leave large quantities of water imprisoned behind it. After the retreat of the wave the imprisoned water escapes through occasional breaches in the barrier and flows down the beach in streams of considerable strength, which scour away the beach material along their courses. The residual masses of material thus left between the stream lines are gradually shaped by the waves into typical beach cusps. A stony barrier would probably not operate in the same manner as a barrier of seaweed, since the water would filter through the mass rather than wear channels. " It would seem to follow that such stony cusps are to be looked for only on coasts where seaweed or some similar material is abundantly thrown up[6]."

In 1900 Branner published a paper entitled " The Origin of Beach Cusps," based on observations made on the California coast and the northeast coast of Brazil. He noted the fact that cusps occur where " there are no seaweeds or other ' drift ' on the beach," and concluded that they are formed " by the interference of two sets of waves of translation upon the beach." The accompanying diagrams, reproduced from Branner's paper,

will serve to make his theory clear. In Figure 139 "the concentric lines represent two sets of waves advancing on the beach in the direction indicated by the arrows and crossing each other along the broken lines. In deep water these are waves of oscillation, but when they reach the shallow water on the beach they become waves of translation and interfere with each other where they converge upon the shore. The tendency is for them to check

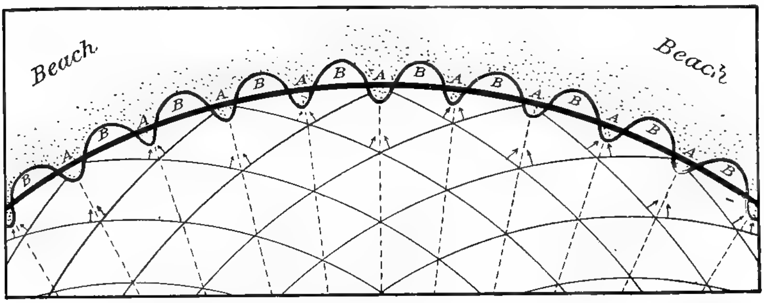


FIG. 139. — Diagram illustrating Branner's theory of beach cusp formation.

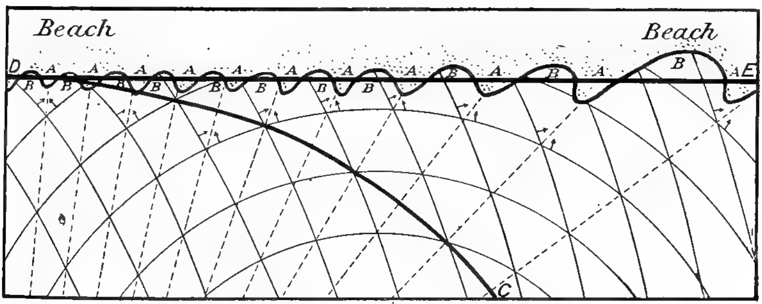


FIG. 140. — Diagram illustrating Branner's theory of the formation of unequally spaced beach cusps. If *DC* were the beach, the cusps would be uniformly spaced.

each other along these lines of interference and to heap up the sands at the points marked *A*, where they strike the beach. At the points marked *B* the waves diverge and throw the beach sands and all floating material alternately right and left."

"In Figure 140 the waves are represented as breaking on a straight beach. If the water offshore were of a uniform depth and the waves were evenly spaced, the cusps in this case would, for obvious reasons, be farther and farther apart from left to

right, as shown along the beach *DE*. The distance between the cusps is equal to the spaces measured on the beach between the radii along which the wave interference approaches the shore[7]." In an editorial note in the Journal of Geology for 1901, Branner briefly restated his theory of cusp formation, and called attention to the fact that "giant ripples" and similar beach structures observed in sedimentary rocks may be fossil beach cusps[8].

Among the "Author's abstracts of papers read at the Washington meeting of the American Association for the Advancement of Science, Section E," published in the Journal of Geology for 1903, is an abstract of a paper by Jefferson entitled "Shore Phenomena on Lake Huron." The abstract suggests a modification of the author's views as published four years before; for while in the earlier paper the possibility of a stony barrier's playing the same part in cusp formation as a seaweed barrier is considered and rejected as improbable, in the later paper we read that the cusps are "component features of a *beach ridge*, . . . The ridge . . . has at times been seen and photographed with water caught behind and rushing out at breaks in the line, as with the weed line at Lynn[9]." Whether or not the breaking of water through the barrier is still thought to originate the cusps is not made clear. The cross-waves noted by Branner were observed by Jefferson, but at no place did he find such waves associated with cusp formation.

Alexander Agassiz in a report on "The Coral Reefs of the Tropical Pacific[10]," figures a series of "boulder cusps" observed on the shores of Arhno atoll. Judging from the illustration these are true beach cusps; but the method of origin advocated by Agassiz is that described on an earlier page of the present volume for the formation of cobblestone deltas in marshes or lagoons by waves washing over a low beach. The position of the "boulder cusps" on the shores of a narrow lagoon, is compatible with the delta theory rather than with the beach cusp theory; but the forms as figured could not have been produced by overwashing waves. Some doubt must therefore attach to Agassiz's brief observations.

In his paper "Cuspate Forelands along the Bay of Quinte"[11] A. W. G. Wilson describes the occurrence of "cusplets" on one of the forelands, and ascribes them to the action of a single

series of waves striking the beach at an oblique angle. Although Wilson does not refer to the previously published accounts, and although the very asymmetrical forms described by him differ in some respects from the essentially symmetrical features generally known as beach cusps, there is little reason to doubt that the former are modified phases of the latter.

In 1905 Jefferson published a paper entitled " On the Lake Shore"[12], in which he gives a brief account of beach cusps, and says " they never occur except after waves that have played squarely on shore." Examples which must have formed without the aid of seaweed barrier are figured, but their origin is not explained. In referring to one particular set, however, Jefferson classes them with the Lynn beach cusps, and says: " Some high wave surmounts the ridge, here of sand, there of seaweed, and its crest water is ponded behind it to escape by any sags that may occur in the line."

My own attention was first directed to the study of beach cusps in the fall of 1903. Seven years later I discussed their form and origin in a paper published in the Bulletin of the Geological Society of America[13], and it is upon this paper that the present discussion of beach cusps is largely based.

Characteristics of Beach Cusps. — When most perfectly developed, the ideal beach cusp has a shape suggesting an isosceles triangle, and is so placed that the unequal side (hereafter called the base) is parallel to, but farthest from, the shoreline. The " triangle " may be short and blunt, or may be so greatly elongated that the two equal sides extend far down the beach and finally unite to form an acute point (hereafter called the apex). These same sides may be relatively straight, but are more often concave, sometimes convex, outward. The actual variations in form are numerous and wide (Fig. 141). Every gradation can be found from well developed triangular accumulations of sand or gravel to widely spaced heaps of cobblestones of no definite shape. The cusps may constitute the serrate seaward side of a prominent beach ridge, or may occur as isolated gravel hillocks separated by fairly uniform spaces of smooth sandy beach. They may be sharply differentiated from the rest of the beach, or may occur as gentle undulations of the same material as the beach proper, and so be scarcely discernible as independent features. Indeed, the variations in beach cusps are so great

Plate LVI.

Photo by James F. Kemp.

Giant sand cusps on Melbourne Beach, Florida, truncated by wave erosion.

that their form is often not as sure a guide to their detection as is their systematic recurrence at fairly uniform intervals. One or two indefinite heaps of gravel on a beach would escape notice, but a hundred such heaps, evenly spaced, attract attention.

A cusp may rise from an inch or less to several feet above the general level of the beach. Many are relatively low and

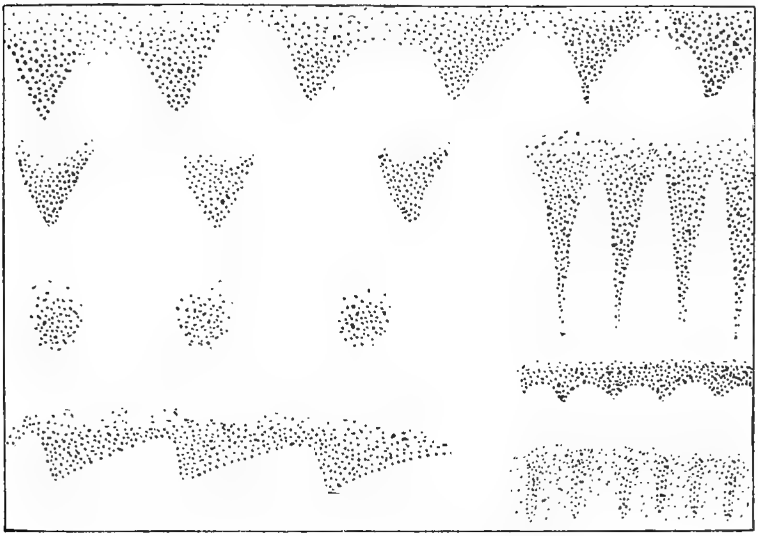


FIG. 141. — Variations in the form of beach cusps.

flat, others high and steep-sided. Sometimes the highest part is comparatively near the apex; at other times the highest part is far back, and from it a long, sloping ridge trails forward toward the water. As a rule, the cusps appear to point straight out toward the water, and neither side of a cusp is steeper than the other except where oblique, wind-made waves have eroded one side only, a condition observed in a few cases.

An interesting variation in form is found where old cusps terminate abruptly in little "cliffs" instead of in sharp points (Plate LVI). It is plain that after the old cusps had been formed they were cliffed by waves under changed conditions and their apices cut away. From this eroded material later series of cusps may form, unrelated in position to the original series. Figure 142 represents a case of this kind as observed in cobblestone

and gravel cusps on a gravel beach at Winthrop, Massachusetts. Sometimes the cusps are more completely eroded than in the case figured, and remnants of three or four distinct sets, of different sizes and spacing, may often be observed on a beach at one time.

As in the form of cusps, so in the material of which they are

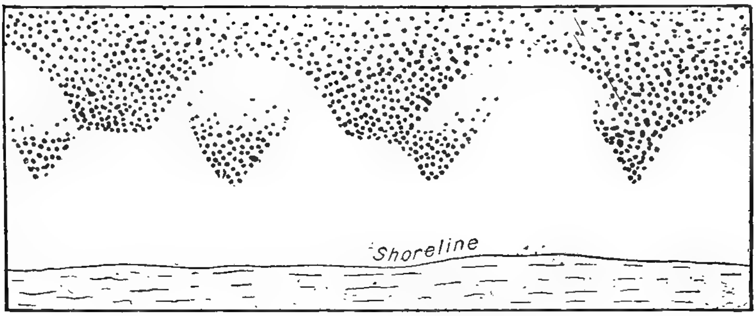


FIG. 112. — Partially eroded older cusps and respaced later series.

composed, is there the widest variation. In building them the waves make use of everything, from the finest sand to the coarsest cobblestones. There is no necessary relation between the size of the cusp and the size of the material of which it is composed. Large cusps built wholly of fine sand are reported from Virginia Beach, and still larger ones (20 to 30 feet from apex to base and 75 to 90 feet between apices) built of similar material were observed on the beach south of Dyker Heights on Long Island. Kemp[14] has studied large sand cusps on Melbourne Beach, Florida, which measured from 90 to 95 feet between apices and rose at least 3 or 4 feet above the general level of the beach. The largest examples are more often built of coarse gravel or cobblestones, while small ones may be composed of either fine sand or coarse gravel. The very smallest cusps, measuring a few inches in length, consist of fine material only, since the small waves which build them cannot transport coarse gravel or cobblestones. Where both coarse and fine materials occur on a beach, the cusps are built of the coarse material. Gravel cusps on a sandy beach are of common occurrence, but I have not observed sand cusps on a gravel beach.

The smallest cusps which have come under my observation

have been those artificially produced in the laboratory. These have varied from an inch to several inches in length, measured from apex to base. Some almost as small are to be found along the shores of sheltered ponds. On a sandy beach at the head of a protected bay south of Huletts Landing, Lake George, cusps from 8 to 12 inches long were formed by the small waves set in motion by a gentle breeze. Those found along the seashore may reach a length of 30 feet or more. It should be noted, however, that the length measured from apex to base is less significant than the distance between cusps, measured from apex to apex; for while it is a general rule that the farther apart the cusps the larger is their size, some which are closely spaced may be greatly elongated, as pointed out above, and this elongation appears to be the result of rather accidental conditions, and to have no great significance. Measurements across the bases might be more significant, but it is often difficult to determine the length of base, as when the cusps form part of a beach ridge or constitute widely separated heaps of gravel having a vague shoreward boundary. However, enough has been said to give some idea of the range in size; and although size is in some degree related to spacing, the latter is the really important factor, as noted below.

The very small cusps made in the laboratory are from one to several inches apart, measured from apex to apex. On the shore of small ponds and bays, where only small waves are developed, the spacing varies from less than a foot to two feet or more. On sea beaches the cusps built by small waves may be less than 10 feet apart, while those built by large storm waves may be 100 feet apart.

Jefferson emphasizes the lack of regularity in the spacing of cusps, whereas others have been impressed by their regular recurrence at fairly uniform intervals. Inasmuch as the matter of spacing is of vital importance in any discussion of the origin of these forms, we may examine it somewhat carefully. Jefferson[15] writes: " The constant *recurrence* of bay (intercusp space) and point (apex) as one walks along the beach suggests that there is a regularity in the width of intervals. This is not so, however, on Lynn Beach, as appears from the diagram, measures from point to point along the beach being 21, 20, 18, 16, 22, 17, 6, 7, and 22 paces. Fainter cusps farther south toward

PLATE LVII.

Cusps on Nantasket Beach, Massachusetts.

Nahant show similar irregularity. It might be said, however, that on Lynn Beach they are commonly about 20 paces wide." And again[16]: " In a view along the beach these unevennesses are foreshortened into the appearance of points of sand or gravel known as beach cusps. They are less even than they look." In still another connection he says: "Perspective foreshortening gives them a fictitious appearance of regularity[17]." On the other hand, Shaler[18] speaks of their " orderly and uniform succession "; and it has seemed to me that the degree of regularity in spacing is so great as to be incompatible with certain of the proposed theories of origin.

It is true that measurements of the spaces do not always give exactly the same figure; that in the early stages of development a greater degree of irregularity prevails than later on; and that even where cusps are very perfectly developed, occasional aberrant features obscure the regularity of spacing. Nevertheless, a large number of observations of beach cusps in all stages of formation and destruction, and the production of artificial cusps in the laboratory have convinced me that a fairly high degree of regularity in spacing is a most characteristic feature of well developed forms and must carefully be considered in any attempt to account for their origin.

The width of the intercusp spaces varies with the size of the waves. When the waves are about an inch in height the cusps are from 3 to 9 inches apart; when the waves are from one and a half to two and a half feet high they are 30 to 60 feet apart, while large storm waves build cusps 100 feet or more apart. These figures are only approximate, and are based on rough estimates of the wave height close to the shoreline. Sufficient data have not been secured on which to base a reliable determination of the precise relation of intercusp space to wave height, but within certain limits there is a suggestion that doubling the wave height doubles the length of the space. A large number of careful observations would probably establish this point. In conducting such an investigation the observer must satisfy himself that the waves he sees are actually building the cusps, for waves of any size may play about cusps formed by other waves of different size, and thus mislead one who compares the intercusp spaces with the height of the later waves. Fortunately a given set of waves does not long leave

unmolested a series of cusps formed by waves of an entirely different size, and the patient observer can in time determine whether or not the waves then breaking on the beach are to be correlated with the cusps at the water's edge.

This brings us to the consideration of another significant point in connection with the spacing of beach cusps: namely, the relative ease with which old cusps are remodeled by waves differing in size from those which formed them. If closely spaced cusps formed by small waves are attacked by larger waves, there ensues a rearrangement by which the cusps become larger and farther apart. This rearrangement may be gradual, and may be accompanied by the combining of some cusps and the slow obliteration of others; or if the new waves are very large, there may be a rapid obliteration of the earlier series of cusps, followed by the slow formation of a new series adjusted to the size of the later waves. If the widely spaced cusps formed by large waves are attacked by smaller waves, so much of the older cusps as can be reached will be eroded and the material refashioned into smaller cusps more closely spaced, regardless of the positions of the older ones (Fig. 142). When large and widely spaced cusps are built by high storm waves well up the slope of the beach, only their apices are apt to be attacked by the smaller waves of calmer weather, and so it happens that we commonly find the largest cusps partially preserved near the top of the beach, with series of smaller and more closely spaced cusps farther down the slope.

Regarding the building of beach cusps, Jefferson[19] writes: "If it be asked how this begins, the answer must be that the beginning is as old as the beach. . . . Each set of cusps may modify its successors. A new crest of seaweed flung up today is likely to have its weak points in some measure determined by the previous channels. In violent storms it is doubtful if this control is significant. Each storm probably sets the shape in which the waves must play for a long time." If we accept Jefferson's theory of cusp formation, the conclusions just quoted would seem to be reasonable. But the sensitiveness of beach cusps to changes in size of waves leads to quite opposite conclusions. Instead of the beginning of cusp formation dating back indefinitely, there appears to be a new and quite independent beginning with every marked change in the size of

waves. One set of cusps seems to have little influence on the position of its successors. Along the shores of a little bay just south of Huletts Landing, Lake George, cusps built by small waves are completely obliterated each day by three or four of the large waves which strike the beach after the passing of a steamboat. Opposite the cusps, but farther up the beach, pegs were driven to mark the position of the cusps. After their obliteration they formed again under the influence of the small waves, with the same size and spacing as before, but, as shown by the pegs, in totally new positions. The law controlling the relation of spacing to wave size was operative, but the cusps which were there a few moments before did not determine the position of their successors. The same phenomenon may be observed in the production of artificial cusps. Furthermore, if a series of parallel trenches be excavated in the artificial beach at right angles to the shoreline, the intercusp spaces and the cusps will not correspond with the trenches and intervening ridges which have been made to guide wave action. In fact, waves of a given size insist on forming cusps at appropriate intervals, and while their action may be influenced within certain limits by natural or artificial trenches on the beach, they refuse to be controlled by such depressions unless these are themselves appropriately spaced. Kemp[20] reports that at Melbourne Beach on the Florida coast continuous observations throughout one winter show that the cusps of one day may be completely obliterated in a few hours, and the beach left featureless and smooth. The next series of waves will form a new series of cusps quite unrelated in spacing to the earlier series.

The bases of the cusps often merge with the last formed beach ridge in such a manner as to leave no doubt that they constitute an integral part of it. The ridge may or may not be breached opposite the intercusp spaces; but it should be noted that with the progressive concentration of the water in the intercusp spaces, which converge shoreward, the parts of the ridge most likely to be broken through are the parts opposite these spaces. It is, therefore, not necessary to regard the intercusp spaces as the product of erosion by water which was imprisoned back of the ridge and broke through it, either at the lowest places or at points of weakness. Conclusive evidence that the ridge may be breached from the seaward side is found

PLATE LVIII.

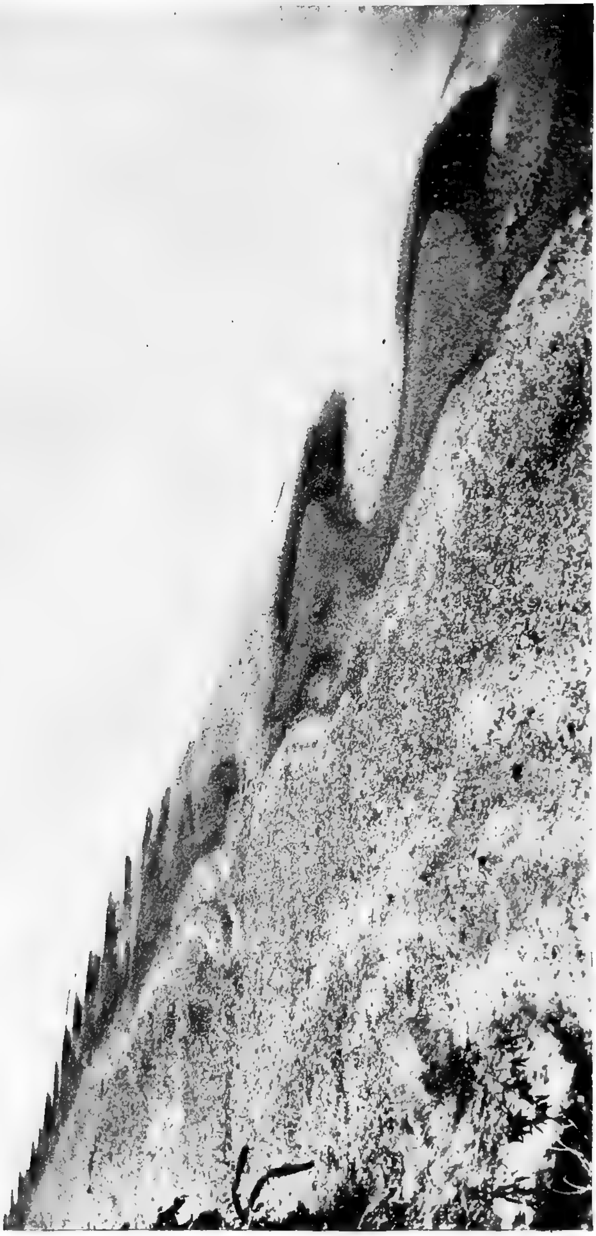


Photo by G. J. Mitchell.

Beach cusps on the west coast of Porto Rico, near Melones Point.

in the gravel or cobblestone deltas which are sometimes built *landward* from the gap in a ridge at the head of an intercusp space (Fig. 143). It seems clear that the water concentrated between cusps broke through the ridge and carried gravel and cobbles into the area back of it. In one case observed at Nahant the landward projection of cobblestone accumulations was so systematic as to give a series of "inverted cusps" alternating

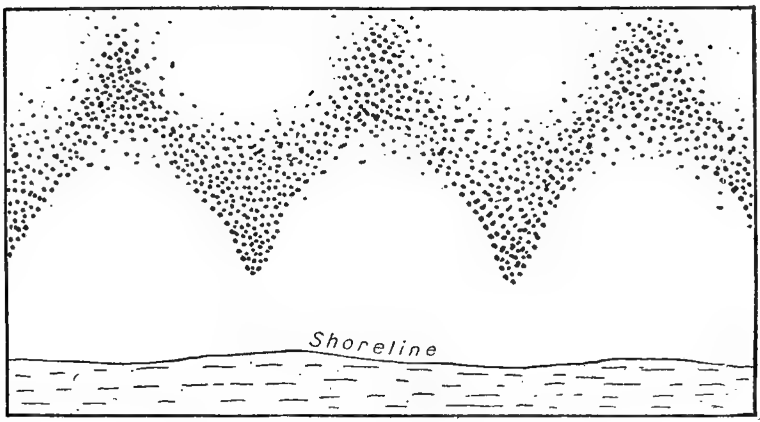


FIG. 143. — Normal and inverted beach cusps.

regularly with the beach cusps proper. The breaching of the ridge by water concentrated between previously formed cusps has been repeatedly observed in the laboratory experiments.

There are abundant instances of cusps unrelated to any beach ridge. Cusps of gravel are often formed at widely separated intervals with smooth, sandy beach between; the points of old cusps are nipped off and respaced without the development of a ridge. One must conclude that cusps may develop as the serrate seaward margin of a beach ridge and may determine the places where it will be breached by the waves, but that there is no necessary relation between the two.

The return current of water flowing down the beach after the wave has ended its advance, sweeps seaward more or less fine material which is deposited to form the shoreface terrace. When cusps have not formed, the margin of this terrace is relatively straight; but after cusps have developed, the greatest

amount of water and débris returns down the slope from the intercusp spaces, building the subaqueous platform seaward more rapidly than does the smaller amount of water and débris returning from around the apices of the cusps. In this way the margin of the platform becomes scalloped, each intercusp space having a scallop or miniature delta to correspond with it. It is evident that the scalloping of the platform presents no difficulty if the origin of the cusps is understood.

Relation of Beach Cusps to other Factors of Shore Activity. — In collecting data concerning beach cusps some attention has been given to several other factors of shore activity, in view of the possibility that they might exert some influence on cusp formation. Several of these factors are briefly treated below.

It was thought at first that the angle of beach slope might exert an important control over the spacing of the cusps, inasmuch as the slope affects both the volume and velocity of the water advancing and retreating over the zone of wave attack. It soon became apparent, however, that if the inclination of the beach does influence the spacing, the effect is largely masked by the far more important factor of wave size. I still think it probable that the slope of the beach plays a small part in the spacing of cusps, but have not sufficient data on this point to demonstrate the truth of the theory.

The direction of the wind seems to have little effect on the formation of cusps. They have been observed in process of formation during onshore, offshore, and longshore winds, both gentle and fairly strong. Under ordinary conditions the only result noticed was a more or less marked cliffing on one side of the cusps when the wind produced small waves at an angle oblique to the beach. The cusps thus cliffed may have been partially developed before the oblique waves began their work. If the wind is strong enough and from such a direction as to combine with the breakers in producing a very irregular wave attack, the formation of cusps is probably interferred with, since numerous observations tend to show that a fairly regular advance and retreat of the water is essential to their development.

Beach cusps are formed at all stages of the tide. It is probable that the greatly elongated type is produced when the waves remain of approximately the same size during a falling tide, but

the development of this type has not been observed throughout the entire process.

The direction of wave advance has been carefully noted whereever cusps were being formed. On the basis of numerous observations on all kinds of beaches and of extended experimentation, it may be confidently stated that the best conditions for cusp formation exist when a single series of waves advances parallel with the beach. It is possible that cusps may be produced by waves striking the shore at a markedly oblique angle, but no satisfactory evidence that such is the case has been secured. On the other hand, the progressive destruction of cusps by oblique waves has been repeatedly observed. Such partially destroyed forms are shown in the lower left-hand corner of Figure 141. I am inclined to think that the asymmetrical "cusplets" reported by Wilson[21] were formerly symmetrical beach cusps of the ordinary type, which were later cliffed by the oblique waves shown in a photograph reproduced in his paper. Intersecting waves of the type appealed to by Branner have been seen in a number of cases, but no cusps have been observed to develop under the action of such waves.

The periodicity of the waves does not appear to be a significant factor in beach cusp formation. Varying the period with artificial waves produces no apparent effect on the cusps.

Jefferson[22] says: "The cusps seem related to a longshore current, their precise cause not being evident"; but he does not indicate in what manner the cusps seemed related to the current. In most of my observations no evidence of a longshore movement of the water was found. In the few cases where a distinct drift or current in one direction was apparent there seemed to be no relation between the current and the cusps. Beach cusps seem clearly to be the product of on- and offshore movements of the water.

Artificial Beach Cusps. — From the observation of natural beach cusps in process of formation the conclusion was reached that cusps could be formed by a single series of waves advancing parallel with the shore. In order to test the validity of this conclusion the artificial production of cusps was attempted. A sand beach was constructed along one side of a tank 5 feet square and the water in the tank raised until it rested against the beach slope. To make that slope as smooth and gentle as

possible, large waves were washed over the beach until it appeared to the eye as a perfectly uniform, gentle slope of sand. On the opposite side of the tank from the beach was arranged the wave-producing apparatus. This consisted at first of a board which was tipped up and down by hand; later of two boards hinged together, one of which was made stationary on the floor of the tank, while the other could be raised and lowered by a long handle connecting with its free edge. With this simple apparatus it was possible to propel on the beach a series of parallel, straight waves, varying in size and periodicity as the experimenter desired. It was found that beach cusps resembling closely those in nature could be artificially produced (Fig. 144). The characteristic features of these artificial cusps have been discussed above.

Fig. 144. — Artificial beach cusps.

Theories of Origin. — With the characteristics of beach cusps in mind, we may critically examine the theories which have been proposed to account for their origin.

The unpublished manuscript of 1887, in which Lane discusses the characteristics of beach cusps, does not set forth a complete theory of their origin, but does contain exceptionally good observations on the more significant features of their occurrence. It will presently appear that some of the significant relationships noted by Lane, and quoted on an earlier page, are necessarily involved in the theory of origin advanced by the present writer.

According to Shaler, "the inner edge of the swash line . . . has a very indented front, due to the fact that it is shaped by a criss-cross action of many different waves[32]." The projecting tongues of water push back the pebbles, leaving indentations or bays, which are then enlarged under the continued wave attack during the rising tide. It should be noted, however, that the indentations of the inner edge of the swash line on a

smooth beach are extremely irregular, and vary in position with every wave advance until the development of cusps and intercusp depressions affords more definite guidance. That a single advance of the irregular inner edge of the swash could develop bays which would thereafter control the action of the waves seems doubtful. The inner edge of the swash is thin as well as irregular and variable, and under these conditions must be very ineffective in developing intercusp spaces or "bays." Nor does the theory as stated by its author explain the regularity in spacing of the cusps nor their respacing consequent upon a change in size of waves. It would seem that Shaler's theory does not go far enough adequately to explain the observed phenomena.

In the account of "ridges and furrows" (cusps and intercusp spaces) given by Cornish[24] it is stated that the water washes depressions at selected places because neither the force of the water nor the resistance of the beach material to erosion is absolutely uniform. The regular spacing of the cusps is not explained, nor does the author appear to have recognized this character of their distribution. Neither does he recognize the fact that gentle waves build cusps. The erosion which produces the "furrowing" is related by him to a change from small to large waves only. But we have seen that cusps form under reverse conditions as well. It thus appears that Cornish points out certain causes of the unequal erosion of beaches, but does not throw much light upon the origin of the cusps.

The seaweed barrier theory of Jefferson[25] advanced to account for the occurrence of cusps on a beach where there happened to be considerable accumulations of seaweed at the time, breaks down under the test of a broader application. There are also serious objections to the theory aside from the fact that cusps are abundantly developed on beaches free from seaweed and other similar material. Even if we admit that a strip of seaweed might form an effective dam behind which considerable masses of water would be imprisoned, we must regard it as in the highest degree improbable that this water would break through the seaweed barrier at a large number of rather evenly and often closely spaced intervals. The degree of regularity in beach cusp spacing is wholly incompatible with the seaweed barrier theory.

On the other hand it should be remembered that after the

cusps have once formed, a seaweed barrier, as well as a barrier of sand or gravel, may be breached by the waves where their water is concentrated for the attack in the intercusp spaces. Thus an observer might find breaches in the barrier corresponding with the intercusp spaces. As shown more fully on a preceding page, both theoretical considerations and the field evidence support the view that the breaching is effected by direct wave attack, and not by the escape of water imprisoned behind the barrier. There is good ground for the belief that the breaching of the seaweed barrier on Lynn Beach was the effect instead of the cause of cusp formation.

In Jefferson's more recent accounts[26] the question of origin is very briefly referred to; but from such reference it appears that the author later considered a barrier of sand or gravel capable of playing the same rôle in cusp formation as a seaweed barrier. It is further implied that other cusps must have had a different but unknown origin. The objections urged against the seaweed barrier theory apply, in the main, with equal force against the sand or gravel barrier theory. It is true that ridges of sand and gravel are more frequent on beaches than barriers of seaweed; but the evidence is conclusive that cusps are formed when such ridges are absent, and that even when present such ridges are breached from the seaward side by direct wave attack, and not from the landward side by impounded waters.

On both natural and artificial beaches more or less distinct ridges are sometimes broken through before any distinct cusps have been formed. This led me to entertain the hypothesis that direct wave attack on a fairly uniform ridge would develop breaches in the ridge at intervals proportional to the size of the waves. It seems probable, however, that faint undulations in the beach, on the seaward side of the ridge, may help to determine the points of breaking just as the more evident cusps and intercusp spaces do in other cases, and that the breached ridges are therefore but one phase, and not an essential one, of the process of cusp formation, as explained on a later page.

Branner's theory[27], while very suggestive, seems to present insuperable obstacles, as will be apparent on the inspection of his diagrams (Figs. 139 and 140). The hypothetical wave lines are evenly spaced, and the wave length in both sets is the same. This is a condition which probably never obtains in

nature, and yet such an improbable condition is an essential element of the theory. If the two sets of waves are given different wave lengths, or if one set of waves has a velocity differing from that of the other, or if either set of waves is irregularly spaced, then the points of wave interference will not reach the beach at the same place twice in succession. If we endeavor to approximate natural conditions by introducing any one of the three types of irregularities mentioned (and probably all three exist in every case of intersecting waves), we must correct the diagrams by making the dotted lines meet the shoreline at every conceivable point. This done, the supposed reason for cusp formation disappears.

It has been shown on preceding pages that the physical conditions necessary for cusp formation exist in parallel waves. One might accordingly surmise that in intersecting waves the necessary equilibrium would be destroyed and the formation of cusps rendered more difficult, or even impossible. I believe this to be the case. In 1907, while camping near Huletts Landing, opportunity was afforded to make numerous observations during a period of six weeks, on a portion of the lake shore where intersecting waves were usually developed by a sand and gravel bar offshore. At no time were cusps observed on the portion of the beach where intersecting waves arrived, although they were frequently found on adjacent portions. These observations led to the belief that intersecting waves tend to prevent rather than to cause the formation of beach cusps.

Inasmuch as the "cusplets" described by Wilson[28] appear to be true beach cusps of somewhat unusual form, it is proper to consider the hypothesis offered to account for their origin. According to this author, evenly spaced waves striking a straight shoreline at an oblique angle will give evenly spaced points of wave-breaking at which cusps will develop. Because at any given instant a series of oblique waves will be breaking at a number of different points along a beach, the author assumes that the points of simultaneous wave-breaking will be nodal points where material will tend to accumulate. It would appear that no account is taken of the fact that every oblique wave of the series breaks not only at the point observed during a given instant, but also at all the other points up and down the

beach, so long as the wave exists. The point of breaking of an oblique wave sweeps along the shore until the end of the wave itself is reached. In a series of waves parallel to each other, but oblique to the shoreline, each wave in turn breaks continuously from one end of the beach to the other. Under these conditions no nodal points can develop, and the fact that the waves are a given distance apart, and that at any given instant their points of contact with the shore are evenly spaced, is immaterial so far as the distribution of force of wave attack is concerned.

In addition to the theoretical objections to Wilson's theory must be added the observed fact that oblique waves appear to be much less favorable to cusp formation than are waves parallel to the shoreline. Oblique waves have been observed in the process of cliffing the sides of cusps exposed to their attack, and the remains of the cusps then have the asymmetrical form decribed by this author.

In attempting to explain the formation of beach cusps I have tested and rejected several working hypotheses in addition to those mentioned above. For example, there was considered the possibility that the waves breaking parallel with the shore had superposed obliquely upon them smaller waves, and that the portions of the main waves thus increased in height excavated the intercusp spaces. One bit of evidence which appeared to harmonize with this theory was personally reported to me by Mr. T. I. Read, who noted that on Virginia Beach the incoming waves showed the first tendency to break at regularly spaced intervals which corresponded with the intervals between cusps. The hypothesis was rejected because the cause was irregular, while the effect was regular; because of an almost complete lack of direct evidence pointing to a relation between superposed waves and cusps; and because experiments seemed to point conclusively to some other origin.

Another hypothesis was based on the assumption that an extended sheet of water descending an inclined plane may not move with the same velocity throughout, but may tend to develop lines of swifter flow, or currents, at certain intervals. I was tempted to make this assumption because of the fact that water descending a flat-bottomed inclined trough, or conduit, does not flow uniformly, but is successively retarded in such a

manner as to produce a succession of waves. Admirable illustrations of this phenomenon have been published by Cornish[29] in a paper on "Progressive Waves in Rivers." It occurred to me that if a broader sheet of fluid were retarded by friction while descending an inclined plane, the resistance might be overcome first, or more rapidly, at certain points, and that the slightly increased rate of advance at these points would disturb the equilibrium in such manner as to create zones or currents of accelerated flow wherever these slight initial advantages had been gained. If the sheet of water were shallow, there would be a tendency for the currents to be smaller and more closely spaced than if the sheet of water were of greater depth. This hypothesis was especially tempting, inasmuch as granting the basal assumption all the phenomena of beach cusps find a ready explanation. Small waves advancing and retreating on the beach would give small currents closely spaced, which would in turn scour small intercusp spaces leaving closely spaced cusps. Any change in the size of waves resulting in a change in the size and spacing of the currents would necessitate a respacing of the cusps. The hypothesis does not lack support so far as the phenomena of beach cusps are concerned, but it is based on an assumption which does lack support. I have questioned a number of engineers and physicists in regard to the matter, but could learn nothing favorable to the assumption.

The hypothesis which best accords with all of the available evidence may now be set forth. Concisely stated, it is that selective erosion by the swash develops from initial irregular depressions in the beach shallow troughs of approximately uniform breadth, whose ultimate size is proportional to the size of the waves, and determines the relatively uniform spacing of the cusps which develop on the inter-trough elevations. This theory differs essentially from those proposed by Branner and Wilson in that neither intersecting nor oblique waves are appealed to and the spacing of the waves is disregarded; from those proposed by Jefferson and Cornish in that the cusps are not regarded as mere erosion remnants of a once continuous ridge, while uniformity of spacing depending on wave size is considered of vital importance; from the theory proposed by Shaler in that no importance is attached to the irregular front of the swash, the ability of the thin edge of the swash to develop

the intercusp bays is not admitted, while the size of the wave is correlated with the width of intercusp spaces. Other points of difference will appear in the explanation which follows.

Every beach contains numerous inequalities which tend to prevent a uniform flow of water up and down the beach during wave action. These inequalities have a variety of causes. Surface run-off after rains may develop channels on the beach; the water draining out of the sand at the upper part of the beach after high tides or after high waves may produce the same result. Pebbles lying on a sandy beach interfere with the swash of water up and down the beach, and cause some channeling. The waves are never even-crested, and may be very irregular if oblique waves are superposed on them; the irregularity of the swash line, mentioned by Shaler, may initiate irregularities on the beach. Remnants of old beach cusps, not wholly obliterated, form another source of irregularity; and still other sources might be mentioned.

The continual swashing of the water up and down the beach tends to enlarge the irregular depressions over which the water passes. Larger channels are better adapted to the movements of the large volumes of wave-supplied water. It is inevitable that in the enlarging of some depressions others will be obliterated, just as in the case of growing drainage basins many small basins disappear as independent features, while the few increase in size. Those depressions on the beach which develop to larger proportions will be the ones which have some initial accidental advantage, and which increase that advantage as they grow; just as the accidentally favored drainage basins increase in size and advantage at the expense of those which began the contest with but a slightly less favorable chance. The tendency of wave action will be to develop from initial irregularities a smaller number of broad and shallow depressions on that portion of the beach traversed by the swash. The depressions will be broad, because they are thus better adapted to the movements of large volumes of water; and shallow, because the elevations between the depressions are also buried under the advancing and retreating waters and are kept worn down to a moderate height. Only near the upper zone of wave action, where the water invades the depressions but does not rise high enough to override the intervening elevations, are the depres-

sions continually scoured deeper and the unworn elevations left as pronounced ridges. Out toward the seaward margin of the submarine terrace, deposition rather than erosion prevails, and the delta scallops may rise higher than the seaward extension of the elevations which exist farther up the beach.

There is a limit to the width to which the depressions, or shallow "channels," if we may so call them, can develop. Inasmuch as the enlargement of some necessitates the obliteration of others, enlargement will continue only so long as the impulse toward growth imposed on the more favored channels is sufficiently great to overcome the tendency of their neighbors to enlarge. Equilibrium will be established when adjacent channels are of approximately the same size, and at the same time of a size appropriate to the volumes of water traversing them. If the waves are low and the volumes of water consequently inconsiderable, equilibrium will be reached while the channels are yet small. But if the waves are high and the volumes of water large, a perfect adjustment will not be reached until the channels have attained great size.

The remainder of the process is easily understood. With the water advancing repeatedly up a beach which is faintly but systematically channeled, as above indicated, there will be a constant tendency to push gravel and other débris farther up the slope in the depressed areas than in the intervening areas. Near the upper limit of wave action the depressed areas alone are invaded by water and are scoured deeper as the gravels are pushed back and the finer material dragged down to form the delta scallops. The intervening areas are fashioned into beach cusps, whose sharpened points divide the waters of the advancing waves and concentrate the attack toward the heads of the depressions. The coarse material is constantly pushed into the cusp areas, the channels swept relatively clean. With a rising tide both channels and cusps are pushed progressively up the beach; with a falling tide some of the gravels may be dragged downward to give much elongated cusps.

There are a number of considerations which appear to support the foregoing theory of beach cusp formation. The theory accounts for the degree of regularity observed in the spacing of beach cusps, since the spacing is dependent on the development of channels which do not reach equilibrium until of approxi-

mately uniform size. At the same time the considerable degree of irregularity in spacing occasionally observed is not incompatible with the theory, since the degree of regularity in spacing depends on the progress which has been made toward the establishment of perfect equilibrium. The occurrence of imperfect and compound cusps is readily explained as the product of wave action in channels not yet eroded to the standard size, as when two unusually small channels have not yet been fashioned into a single large one, and consequently give a compound cusp (Fig. 145) near their upper limits. We should expect, on the basis of this interpretation, that irregular and compound cusps should be most characteristic of the early stages of development, and the

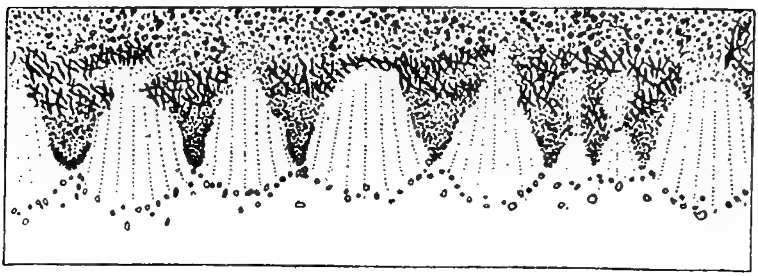


FIG. 145. — Beach cusps (after Jefferson) showing compound cusps at right.

experiments with artificial cusps prove most conclusively that this is the case. One of the commonest occurrences in the experiments is the gradual moulding of irregular and compound cusps into simple cusps regularly spaced.

The respacing of cusps with a change in size of waves may be thus explained: A given set is formed and driven up the beach, and then left by the falling tide. The size of waves changes, and new channels appropriate to them are formed. New cusps result, and as the tide rises these are in turn pushed up the beach. If the new cusps do not coincide in position with the older ones, when the latter are reached their ends will be eroded by the waters converging on them from between the new ones. Repetitions of this process, with waves of decreasing size, will give several sets of partially preserved cusps, each set more closely spaced than the set above it. On the other hand, if a big storm drives in unusually high waves, big channels will be formed,

older sets of cusps will be quickly swept out of existence, and a single set of large, widely spaced cusps will be developed.

In the laboratory experiments difficulty was often experienced in getting the cusps started. The artificial beach was very smooth, of fairly uniform sand grains. It appeared that the difficulty was due to the regularity of the beach, on account of which the initiation of channels was delayed. In order to facilitate the process a series of closely spaced creases down the beach was made, after which the cusps began to form more rapidly. As already shown, the artificial creases did not control the number or position of the cusps and their intervening spaces, but the importance of initial depressions in the cusp-making process seemed clearly indicated.

On Westquage Beach, Rhode Island, the writer has watched a series of parallel "creases," or rill lines, without any associated cusps, develop into channels or intercusp spaces with fairly good associated sand cusps. Such observations are relatively rare, however, probably because the initial irregularities are often indistinct undulations in the beach surface or are soon transformed into such undulations; and because the successive changes in the form of broad, shallow channels on a gravel or sand and gravel beach are difficult to trace. The "ribbed" structure occasionally reported by observers looking for cusps probably represents an early stage of cusp formation.

The tendency of intersecting or criss-cross waves would be continually to shift the sands first in one direction and then in another obliquely over the beach, and thus to prevent the formation of systematic channels. This would account for the observed failure of such waves to form beach cusps, although they might attack cusps previously formed, or leave a beach with irregularities which might affect the formation of later cusps.

In a similar way, to a less extent, a single series of oblique waves would not seem favorable to cusp formation, because of the lateral element in the movement of the water, which would continually tend to wash the interchannel elevations into the channels, and so to fill them up.

It is not necessary to review all the details of beach cusp characteristics in connection with the theory set forth above. It is sufficient to state that the author has found no feature of

beach cusps which is incompatible with the theory, while the assumed conditions of wave action appear to rest on a reasonable basis.

LOW AND BALL

The shoreface zone, or possibly the inner margin of the offshore zone, is frequently characterized by submarine bars or ridges, separated by distinct longitudinal depressions and lying parallel to the shoreline. English writers apply the name *ball* to the ridges and *low* to the depressions. The continuity of the ball is sometimes truly remarkable, Russell[30] describing examples on the shores of Lake Michigan which " can be traced continuously for hundreds of miles." In this case " there are usually two, but occasionally three, distinct sand ridges; the first being about 200 feet from the land, the second 75 or 100 feet beyond the first, and the third, when present, about as far from the second as the second is from the first. Soundings on these ridges show that the first has about 8 feet of water over it, and the second usually about 12; between, the depth is from 10 to 14 feet . . . . They follow all the main curves of the shore, without changing their character or having their continuity broken." Russell suggests that these balls may represent accumulations of shore débris along the lines where the undertow loses its force during storms of varying degrees of intensity; but qualifies the suggestion with the statement that " the complete history of these structures has not been determined."

The balls of Lake Michigan were earlier described by Desor[31], who in 1851 attributed them to transportation and deposition by " currents," and stated his belief that the elevated beaches about the Great Lakes were really submarine bars of the same type which had been exposed to view by a rising of the land. Whittlesey[32] treats them briefly as a product of " lateral currents." In 1870 Andrews[33] called attention to the " subaqueous ridge or bar " which is " uniform in all the sand shores " at the head of Lake Michigan. Gilbert at first[34] considered the balls of the Great Lakes region as barrier beaches or spits built at the lake surface and later submerged by a rise of the waters; but later[35] decided that they were originally formed as subaqueous bars. Concerning the method of their formation

he writes: "Under conditions not yet apparent, and in a manner equally obscure, there is a rhythmic action along a certain zone of the bottom. That zone lies lower than the trough between the greatest storm waves, but the water upon it is violently oscillated by the passing waves. The same water is translated lakeward by the undertow, and the surface water above it is translated landward by the wind, while both move with the shore current parallel to the beach. The rhythm may be assumed to arise from the interaction of the oscillation, the landward current, and the undertow."

The earliest description of low and ball of which I find record is given by Hagen in his "Handbuch der Wasserbaukunst[36]." Hagen considers the phenomenon a normal characteristic of a gently sloping sea-bottom, and refers to a popular belief that three parallel balls ("Riffe") are always found in association. He shows, however, that the number is not constant, as many as five sometimes being revealed by careful soundings. The ridge nearest the shore is highest, those farther out progressively decreasing in altitude until the outermost may rise an almost imperceptible distance above the sea floor. In Hagen's opinion the ridges form where on-coming waves meet the undertow, especially where the undertow is reinforced by backward moving water of normal oscillatory waves.

A brief account of the form of parallel balls is given in Braun's "Entwickelungsgeschichtliche Studien an europäischen Flachlandsküsten und ihren Dünen," under the caption "Das Sandriff[37]." He follows Lehmann[38] in considering the ball as a forerunner of the offshore bar or beach ridge, the ball being driven landward and ultimately raised above sealevel by the action of the waves. Observations of European examples lead to the conclusion that normally the landward side of the ball is steeper than the seaward slope. Otto, on the contrary, in a full description of these submerged ridges published in his work on "Der Darss und Zingst[39]" finds them more variable in form and in behavior. They are sometimes evenly, sometimes irregularly spaced, and often migrate seaward as well as landward. Sudden and marked changes in the ridges occur only with great storms. A comparison of wave lengths and the distances between ridges shows that no correspondence exists between the two measurements. Both Braun and Otto give a short

bibliography of the subject, which should be consulted by those desirous of securing further data regarding the lows and balls of the Baltic shores and other coasts of continental Europe.

Under the title " Low and Ball of a Sandy Shore[40]," Cornish states that the building up of a " full " of sand in front of the breaker is accompanied by the excavation of a trough, at the back of the breaker. Beyond the trough there rises a sandbank which is called the ball, while the trough itself is the low. Ebb tide may reveal the surface of the ball, under which condition a lagoon occupies the low between the ball and the beach. Wheeler[41] also speaks of the low as a gully running parallel to the coast cut by the action of breakers, and is of the opinion that the ball may rise permanently above the water surface, causing a permanent lagoon or shallow creek in the adjacent low.

Kemp[42] has recorded some valuable observations regarding the lows and balls of the Florida east coast. At Melbourne Beach, and for an indefinite distance north and south, the shore is normally bordered by a distinct channel varying in breadth from 15 to 60 yards and usually not so deep but that bathers could walk across it to the bar beyond at low tide. The crest of the bar rose within a few inches of the water surface, but was never seen exposed. Those engaged in surf-fishing for "channel bass" become familiar with all changes in the low, for this is the channel in which the bass run. After maintaining a fairly constant position for three months in the winter of 1915–16, the bar migrated shoreward under the influence of heavy surf from a strong easterly gale. After the storm died down the bar continued its shoreward progress until the low was reduced to a breadth of 5 yards, then 2 yards, and finally was extinguished. The next fall the fishermen found a new bar with broad channel intervening between it and the shore, just as at the beginning of the previous winter.

In classifying the forms observed on the Great Lakes by Desor, Gilbert, Russell, and others, with those observed on tidal shores by Cornish and Wheeler, and giving the English names low and ball to the entire series, I have proceeded on the assumption that the two forms are similar in character and identical in origin, such differences as are noted being due to the changing water level in the case of the marine type. I must state, however, that this procedure is not based on any careful

comparison of these forms as developed in lakes and in the ocean, and my classification is accordingly to be accepted with due reservation. While I have examined fairly good lows and balls along the sandy beach at Cape Henry, Virginia, and elsewhere on the Atlantic shoreline, I have not seen those of the Great Lakes; nor have I made sufficient study of the examples observed to add anything of value to the discussion of their origin.

RIPPLE MARKS

The accumulation of sand and finer débris in parallel ridges and troughs somewhat resembling water waves in form, though not at all in origin or method of formation, was long ago recognized as a normal product of wave and current action. Under various names, such as " current mark," " wave mark," " ripple drift," " current drift," and " friction markings," the phenomenon now generally known as *ripple mark* has repeatedly been described. Although not infrequently found on sandy beaches, ripple marks are perhaps better developed on tidal flats and over the broad shallow bottoms of estuaries. They are not unknown on the deeper sea floor of the offshore zone, where their occurrence to a depth of over 600 feet has been demonstrated. Ripple marks exposed by the falling tide may be delicately dissected by rill marks, an interesting example of this phenomenon having been described by Dodge[43].

Among the earlier accounts of ripple marks one of the most interesting is based on the little known work of an ingenious French engineer named Siau[44]. In 1841 this investigator published a brief note entitled " De l'action des vagues à de grandes profondeurs," based on observations of ripple marks in deep water made with the aid of an ordinary sounding apparatus. While examining ripple marks, visible during quiet water, on the bed of a channel off the west coast of the Isle of Bourbon, Siau noted that the heavier particles of the sand tended to accumulate in the troughs between the ridges, while lighter material was concentrated along the ridge crests. Profiting by this discovery, he coated a sounding lead with tallow, and lowered it to the sea floor where the depth was too great for direct visual observation. When brought to the surface the tallow sometimes retained, adhering to it, only heavy particles of sand, in

PLATE LIX.

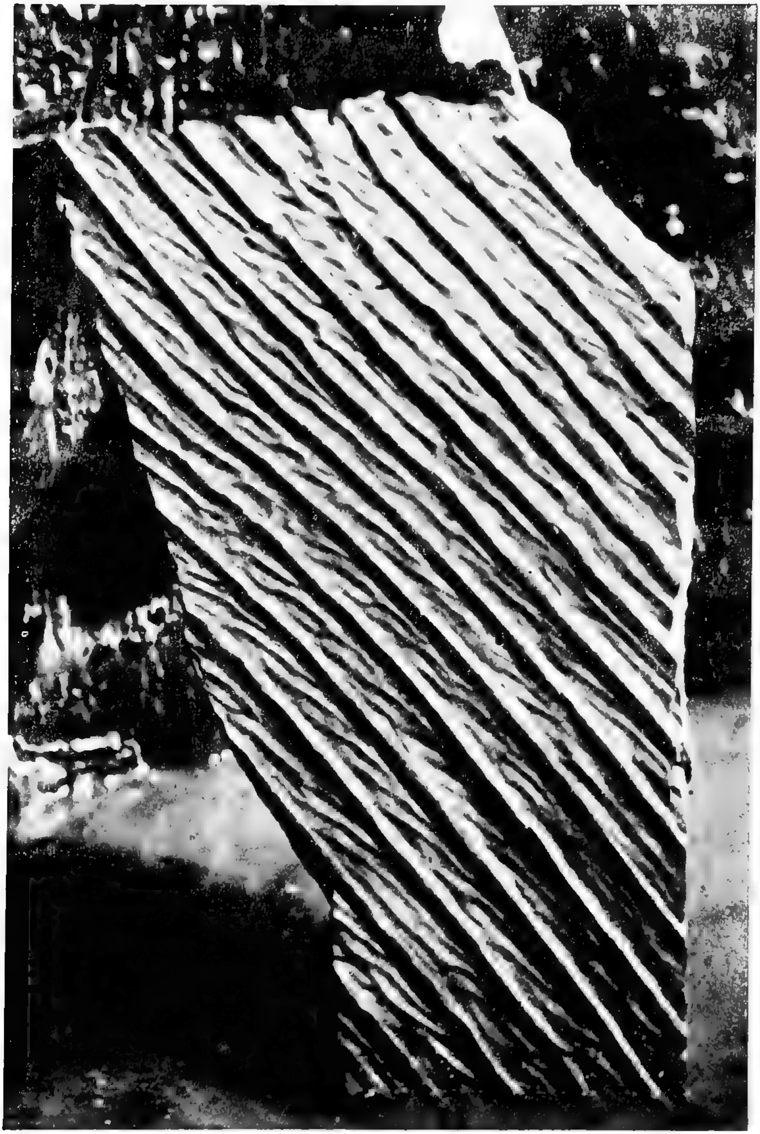


Photo by G. K. Gilbert, U. S. G. S.

Sandstone slab showing fossil oscillation ripples. A later, smaller series of oscillation ripples had begun to form in the troughs of the main series.

which the surface of the tallow had convex form, showing that it had been pressed down into the trough between two ripples. In other cases the tallow was coated with lighter particles only and had a concave form, as a result of having been pressed down upon a ripple crest. At great depths, where the ripples were more closely spaced, two parallel bands of materials, differing in specific gravity, would be impressed upon the tallow at the same time, the heavier material coating a convex ridge and the lighter a concave depression in the tallow. By this ingenious device Siau was able to prove the existence of ripple marks at a depth of 617 feet.

The ripples described by Siau were believed by him to be due to the back-and-forth currents, which, as we have already seen, are produced on a sea-bottom by oscillatory waves. Such ripple marks are called "oscillation ripples," and are characterized by symmetry of crests, neither slope being steeper

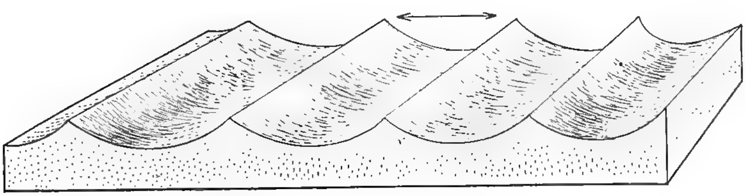


FIG. 146. — Oscillation ripples.

than the other, since the ridges are built up by currents which operate from either side with approximately equal force. The crests are sharp and narrow as compared with the more broadly rounded intervening trough (Fig. 146). De la Beche[45] in his Geological Observer describes another type of ripple mark produced by the action of a current flowing steadily in one direction over a bed of sand. These "current ripples" have a long, gentle slope toward the direction from which the current comes, and a shorter, steeper slope on the lee side. Sand grains removed from the gentle slope are carried to the crest and dropped down the steeper slope, causing the ripples to migrate slowly with the current, much as sand dunes migrate with the wind. The asymmetry of profile of the current ripple is shown by Figure 147, and is apparent in Plates LX and LXI Barrell[46] and others restrict the term "ripple mark" to oscillation ripples,

PLATE LX.

Photo by M. C. Dickerson.

Current ripples formed by an ebbing tide. The current moved from left to right.

and employ the term "current mark" for the asymmetrical type. This usage has much to commend it, but is open to several objections. The fact that "current mark" is produced by water currents might lead to the inference that "ripple mark" is produced by water ripples, which is not at all the case. Ordinary waves rather than true ripples commonly produce

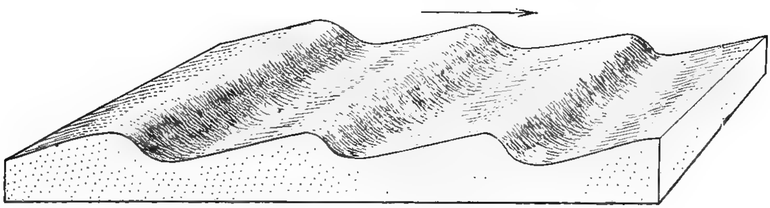


FIG. 147. — Current ripples.

oscillation ripple marks. There are, moreover, other markings produced by currents, as will appear on a later page. The term "ripple mark" is so firmly established in the literature to include both the symmetrical and asymmetrical types that it seems wisest to follow this usage, prefixing the words "oscillation" and "current" to make clear the necessary distinction.

Sorby[47] gave a very good description of current ripples in The Geologist for 1859, but failed to recognize the existence of wave-formed oscillation ripples, although he noted, and even pressed too closely, the analogy between true waves and ripple mark. For many years current-formed ripples were the only type recognized in most textbooks. Gilbert[48] in 1875 described briefly what appear to have been oscillation ripples, but explained them as the product of running water thrown into vibration by friction on the bottom, a theory apparently similar to the "intermittent friction" theory of de Candolle, described below. In 1882, in opposition to the general view, Hunt[49] claimed that as a rule ripple marks are the product of oscillatory wave action, and supported his claim with observations based on the artificial production of ripple marks, as well as with numerous citations of naturally formed ripples. He was evidently unaware of the fact that Siau had supported the same theory some 40 years earlier, and in a later paper[50] erroneously credited Forel with priority in the recognition of oscillation ripples.

Hunt incidentally describes oscillation ripples in his paper " On the Action of Waves on Sea-Beaches and Sea-Bottoms[51] "; he also discusses the nomenclature of ripple marks at much length in a paper published in 1904[52], and elsewhere quotes Lieutenant Damant, R.N., as having observed ripple marks while diving at depths of 60 and 70 feet[53].

In 1883, the year following the publication of Hunt's earliest paper quoted above, there appeared three important essays on ripple marks: one by de Candolle on " Rides Formées à la Surface du Sable Déposé au Fond de l'Eau et autres Phénomènes Analogues"; another by Forel on " Les Rides de Fond Etudiées dans le Lac Léman "; and a third by Darwin " On the Formation of Ripple Mark in Sand." De Candolle[54] produced ripple marks artificially by experimenting not only with sand and various substances in powdered form covered by water, but also with liquids of varying viscosity covered with water and other liquids. Regarding sand or powder when mixed with water as a viscous substance, he concluded from his experiments that " When viscous material in contact with a fluid less viscous than itself is subjected to oscillatory or intermittent friction, resulting either from a movement of the covering fluid or from a movement of the viscous mass itself with respect to the covering fluid, (1) the surface of the viscous substance is ridged perpendicularly to the direction of friction, and (2) the interval between the ridges is directly proportional to the amplitude of the friction-producing movement." That ripple marks depend on simple friction alone, and not on any change of level in the covering liquid such as occurs during wave action, de Candolle proved by an experiment with a rotating disc submerged in a tank of water. After submerging the disc and mixing an insoluble powder in the water, the apparatus was left until the powder settled on the disc and floor of the tank as an even film, and the water came to rest. An oscillatory rotary movement then applied to the disc caused radiating ripples to form upon it, while no ripples formed on the stationary bottom, and the surface of the water remained quiescent. The author concludes that the formation of ripples in sand, whether under currents of air or under water currents, is identical in origin with the formation of water ripples under moving air. If the current moves always in one direction we have intermittent friction due to varying

PLATE LXI.

Photo by F. Berckhemer.

Current ripples on the shore of Nantasket Beach, Massachusetts. The current passed from right to left.

velocities. Otherwise we have oscillatory friction due to alternating change of direction. Current ripples result from the first type of friction, oscillation ripples from the second.

Forel[55] in his excellent essay on " Les Rides de Fond Étudiées dans le Lac Léman " sets forth the mature results of studies which had been briefly mentioned by him in three communications[56] of earlier date. Abandoning his first theory, that the formation of ripple marks was dependent in part upon the vertical pressure of water waves upon the bottom[57] Forel reached the following important conclusions as the result of many careful observations and experiments: (1) Current ripples are asymmetrical and migrate with the current like ordinary sand dunes, whereas oscillation ripples are stationary and symmetrical. (2) Each oscillation ripple is really a composite of two current ripples resulting from the action of two currents moving alternately in opposite directions, each current attempting to form the ridge into a current ripple migrating with it, but being defeated when the return current tries with equal force to shape the ridge into a current ripple directed in the opposite sense. (3) The length of the water body has no direct effect on the spacing of the ripples. (4) Other things equal, the ripples are more closely spaced with increasing depth. (5) At a given depth, and with other conditions uniform, the ripples are more widely spaced with increase in coarseness of sand grains. (6) Ripples once formed do not experience a change in spacing as a result of diminishing amplitude of oscillation of the water, although the original spacing does depend upon the amplitude of oscillation, as pointed out by de Candolle. (7) For any given coarseness of sand grains there is a certain mean velocity of the oscillating currents which will produce ripples: lower velocities will fail to move the sand grains, and hence cannot build ripples, while higher velocities agitate the whole mass of sand so violently that no ripples can form. (8) The formation of ripples is initiated by some obstacle or inequality on the surface of the sand, behind which sand grains accumulate in the eddy caused by its presence: this leaves a furrow on either side of the initial ridge, due to the abstraction of sand accumulated in the ridge; and these furrows in their turn cause additional ridges to develop on their outer margins, and so on. (9) In a given locality, ripple marks almost always form with the same spacing, regardless of

the varying intensity of winds and waves affecting the water body; this is in consequence of laws 7 and 6 stated above. (10) The depth at which ripple marks may form is limited by the depth to which wave action may extend with sufficient energy to move the bottom sands; hence it depends on the size of the waves, and therefore in part indirectly on the size of the water body: in the Rhone the limiting depth is a few decimeters; in Lake Geneva some ten meters; and in the ocean from 20 to 188 meters, according to Lyell and Siau. Forel revised de Candolle's law regarding the relation of ripple spacing to amplitude of the friction-producing movement to read: "The breadth of the ripples, or the distance from one crest to another, is the length of the path followed during a single oscillation by a grain of sand freely transported by the water." The length of this path varies directly as the horizontal amplitude of the oscillatory movement of the water, directly as the velocity of that movement, inversely as the density of the sand, and inversely as the size of the sand grains.

Darwin's paper[58] "On the Formation of Ripple Mark in Sand" is especially noteworthy for its careful analysis of the

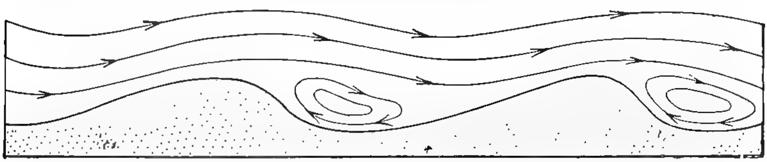


Fig. 148. — Vortices involved in the formation of current ripple mark.

vortices which are so important a factor in the construction of the ripples. When symmetrical oscillation ripples were subjected to the action of a steady current, Darwin noticed that not only did sand grains migrate up the weather slope of each ripple with the current, but that they also ascended the lee slopes, apparently *against* the current. This proved conclusively the existence of such vortices as are represented in Figure 148. Darwin then proceeded to study the vortices by watching the movements of a drop of ink released from the end of a fine glass tube at that point in the water where the action was to be observed. In this manner the vortices associated with the alternating currents which produce oscillation ripples were analyzed

with a high degree of precision, and much light was thrown upon the method of ripple growth. Darwin concluded that "the formation of irregular ripple marks or dunes (current ripples) by a current is due to the vortex which exists on the lee side of any superficial inequality of the bottom; the direct current carries the sand up the weather slope and the vortex up the lee slope. Thus, any existing inequalities are increased and the surface of sand becomes mottled over with irregular dunes." The intermittent friction to which de Candolle appealed is not essential in this explanation of current ripples. Oscillation ripples of regular pattern are changed by a continuous current into regularly spaced current ripples; but a uniform current cannot of itself initiate regularly spaced ripple mark. "Regular ripple mark (oscillation ripples) is formed by water which oscillates relatively to the bottom. A pair of vortices, or in some cases four vortices, are established in the water; each set of vortices corresponds to a single ripple crest." Forel's conception of an oscillation ripple as a composite of two dunes (current ripples) formed alternately by oscillating currents is regarded as correct; but his law for the relation of ripple spacing to amplitude of oscillation is believed to require some modification.

Further studies of ripple-forming vortices were made by Mrs. Hertha Ayrton[59] the results of which were not published until 1910. With the aid of well-soaked grains of ground black pepper, or of particles of potassium permanganate dissolving and coloring the water while the latter was in oscillation, she observed the formation of vortices and endeavored to explain the mechanics of their growth. Although she expressed disagreement with the conclusions of Darwin and others on certain points, most of her results afford essential confirmation of their main contentions. Some doubt must attach to certain of her deductions, such as one to the effect that no ripple-forming vortex occurs in the lee of an obstacle over which a steady current is passing and hence "a steady current is unable either to generate or to maintain ripple-mark."

The British Association Reports for the years 1889, 1890, and 1891, contain three papers by Reynolds[60] on the action of waves and currents in model estuaries, in which are some valuable observations regarding what may well be termed giant tidal ripples. While experimenting with artificial tidal currents

PLATE LXII.

Photo by G. K. Gilbert.

Giant current ripples near Annisquam, Massachusetts, showing irregular pattern due to interference of wave and tidal currents.

Reynolds discovered that current ripples were formed in the model estuaries. By making due allowance for the difference in size between the model estuaries and those in nature, he concluded that real tidal currents ought to produce very large current ripples, possibly 7 or 8 feet in height and 80 to 100 feet apart[61]. Some years later Vaughan Cornish[62] discovered natural tidal ripples or " sand waves " of the same type as those produced artificially by Reynolds, having a height of 2 feet and an average distance of more than 37 feet from crest to crest. In two later papers[63] Cornish describes giant tidal ripples more fully, and illustrates their essential features with a large series of beautiful photographs. Some of these ripples have a height of nearly 3 feet above the intervening troughs, and a distance between crests of from 66 to 88 feet in extreme cases. The giant ripples are often covered with ordinary ripple mark, and while Cornish recognized that the larger forms were produced by the continuous steady flow of tidal currents, he was at first inclined to invoke pulsatory currents in order to explain the smaller ripple mark[64]. This theory seems to be a survival of de Candolle's erroneous idea that " intermittent friction " was essential to the production of current ripples, and is practically abandoned by Cornish in his more recently published book on " Waves of Sand and Snow[65]." Gilmore[66] describes tidal ripples on the Goodwin Sands having a height of " two or three feet," and Kindle[67] reports " mammoth tidal ripples " in estuaries of the Bay of Fundy varying in length from 2 feet up to 15 or 20 feet, and in height from 6 inches to nearly 2 feet. Gilbert[68] measured examples near Annisquam, Massachusetts, which were 15 feet in length and 15 inches high, Plate LXII. River currents as well as tidal currents are capable of forming giant ripples, and Kindle[69] describes examples formed on a broad sandbar in the Ottawa River at time of flood which measured 30 to 45 feet in length and from 1 to 2 feet in height. The same author quotes Pierce as authority for the existence in the San Juan River in Utah of examples rising 3 feet above the adjacent troughs. Unfortunately Pierce[70] improperly applies the term " sand wave " to the water wave formed at the surface of a current passing over true sand waves or giant ripples. It is not clear that Pierce either saw or measured the giant sand ripples, supposed by him to have caused the surface water waves to which his figures apply.

It should be noted that all of the giant ripples referred to above belong to the unsymmetrical type; they are true current ripples. So far as I am aware no giant oscillation ripples have ever been observed along modern shores. It may be doubted whether tidal currents could form symmetrical ripples, notwithstanding Reynold's suggestion to the contrary[71]. The flow and ebb of the tide constitutes an oscillating current, it is true; but the currents are often of unequal force. Where equally strong, each current persists long enough to remodel the ridges formed by the preceding current, giving them an asymmetrical form appropriate to the current operating last. On the other hand, Gilbert[72] has described structures in the Medina sandstone formation of New York which he believes to be giant ripples of the symmetrical type formed by oscillating currents due to wave action. In dimensions these ridges were similar to the average examples of tidal ripples described by Cornish, having a height of from 6 inches to 3 feet and a distance from crest to crest of 10 to 30 feet; but their nearly symmetrical form did not suggest a similar origin. Gilbert reached the tentative conclusion that they were formed by waves 60 feet high in deep water of a broad ocean. This conclusion was criticized by Fairchild[73], who advanced convincing arguments in support of the opinion that the forms in question were beach structures, possibly successive beach ridges built on the strand. Branner[74] suggested that they might represent fossil beach cusps seen in cross-section.

Some interesting experiments on the relation of current velocity to ripple-mark formation were made by Owens[75], who published his results in 1908. He found that currents from 0.85 to 2.5 feet per second produced or maintained a rippled surface on sand; but that a velocity of 2.5 feet per second and above swept the surface free of ripples.

In 1911 A. P. Brown[76] published a paper entitled "The Formation of Ripple-Marks, Tracks and Trails" in which he endeavored to show that asymmetrical ripples (current ripples) were formed by deposition, whereas symmetrical ripples (oscillation ripples) resulted from the erosion of a formerly smooth bottom consequent upon the rippling of overlying water by wind action. His conclusions do not appear to be supported by a sufficient body of convincing evidence, and are opposed

by theoretical considerations and by the great body of experimental data already referred to on previous pages. Unfortunately, in presenting his theory Brown does not consider the important results obtained in the many previous investigations of ripple marks.

A similar criticism must be urged against the work of Epry[77] who in 1912 published a paper on " Les Ripple-Marks " in the Annales de l'Institut Océanographique. Epry states that no one before him has been able accurately to determine the causes of ripple marks and that no previous theory of their origin is satisfactory. He fails, however, to show wherein earlier theories are defective and from his essay it does not appear that he was acquainted with the various publications cited above. Current ripples and oscillation ripples are not distinguished by him, and a highly specialized theory of origin, impossible of application to the majority of ripple surfaces, is developed. It is not necessary to criticize Epry's theory in detail, but a general idea of its essential nature may be gathered from the fact that it involves the remarkable assumption that ripples are formed where an ebbing tidal current returning from the shore is cut transversely by another current deflected along a depression in the sea floor, and that the ripples are aligned in the direction of (parallel to) the transverse current. No less remarkable is Epry's statement that ripple marks are the work of tides alone.

We have already noted that current ripples, like sand dunes, normally migrate slowly in the direction of the current which is fashioning them. Vaughan Cornish[78] discovered, however, that in shallow water when the current attains a velocity of about 2.2 feet per second, the ripples travel upstream or against the current. This observation was later confirmed by Owens[79], and the phenomenon is described by Gilbert[80] in the following words: " When the conditions are such that the bed load is small, the bed is molded into hills, called dunes, which travel downstream. Their mode of advance is like that of eolian dunes, the current eroding their upstream faces and depositing the eroded material on the downstream faces. With any progressive change of conditions tending to increase the load, the dunes eventually disappear and the débris surface becomes smooth. The smooth phase is in turn succeeded by a second

PLATE LXIII.

Photo by E. M. Kindle.

Current ripples near Windsor, Nova Scotia. The current moved from right to left.

rhythmic phase, in which a system of hills travels upstream. These are called anti-dunes, and their movement is accomplished by erosion on the downstream face and deposition on the upstream face. Both rhythms of débris movement are initiated by rhythms of water movement." Pierce[81] states that the anti-dune movement is best seen "only on heavily loaded silt streams," and cites cases of the phenomenon in the San Juan River in Utah.

The best recent essay on ripple marks is a paper by Kindle[82] entitled "Recent and Fossil Ripple Mark," published in 1917. This author presents an excellent summary of his own extensive observations, distinguishes the different types of ripple marks and their methods of origin, and gives many references to the work of others. The abundant illustrations contain some of the best views of ripple marks ever published. Kindle studied different types of ripples by means of plaster casts, some of which were secured at depths ranging up to 27 feet by means of a specially devised apparatus. Siau's experiments were also imitated by lowering to the bottom, at any depth, a rectangular plate of sheet iron or zinc, the under surface of which had been coated with vaseline. Where ripple marks occurred, parallel lines of sand adhering to the vaseline indicated the position and spacing of the ripple crests. On the basis of his studies Kindle concludes that the length of asymmetrical or current ripples varies with the velocity of the current, with the volume of sediment in suspension, and possibly also with depth. "A strong current carrying a maximum load of sand probably forms ripple mark of large amplitude (length) where a slightly loaded current having the same velocity would leave no ripple mark." The author is less certain about the factors controlling symmetrical or oscillation ripples, but thinks coarseness of sand, depth of water, and length of the water waves are of chief importance. In studying Kindle's valuable paper the reader must guard against misapprehension arising from his use of the term "amplitude" to denote the *length* of both sand waves and water waves.

Some of Kindle's conclusions must be regarded as open to question. This is particularly true of the following generalizations: "On the shores of lakes where ripple mark is due entirely to wave action it always runs parallel with the coast-

line. Ripple mark along the sea coast is generally the work of tidal currents which follow the shoreline. These current-made ripple marks consequently trend at right angles to the coast-line. Lake shore and sea shore ripple mark are thus differently oriented with respect to their adjacent shorelines, the former trending with the shoreline, the latter at right angles to it[83]"; "the abundance of the wave-made type of ripple mark in a sandstone formation and the absence of the asymmetrical type would indicate its formation under lacustrine conditions. The great predominance on the other hand of the asymmetrical type of ripple mark would as certainly suggest the work of tidal current action and marine conditions[84]." My own observations of ripple marks do not tend to support the conclusions expressed in these quotations. While it is true that wave refraction often brings about a more or less perfect parallelism between wave crests and the shoreline in the immediate vicinity of the latter, the parallelism is, on the other hand, often far from perfect; and a few feet from the shore the waves not infrequently trend at large angles to the shore. I have, on a number of occasions, observed ripples on the bottoms of ponds and lakes which were, like the waves which formed them, not parallel to the shoreline even when but a few feet distant from it. The supposed restriction of oscillation ripples to lacustrine deposits seems equally doubtful. Some of the best oscillation ripples I have ever seen were formed on tidal flats, bordering the Long Island shore, by wave action when shallow water covered the flats at high tide. Other good examples may frequently be seen in shallow ponds and abandoned channels on river flood plains. Kindle's discriminations between marine and lacustrine deposits (see pp. 48 to 51 of his essay), and between lacustrine and fluviatile deposits (pp. 52 and 53), on the basis of the type of the contained ripple marks, must therefore be accepted with caution, just as truly as must his deductions regarding the direction of ancient shorelines based on the orientation of fossil ripple marks. Even where a geological formation contains ripple marks exhibiting a remarkable uniformity of orientation over wide areas, as in a case described by Hyde[85] in a valuable paper published a few years ago, and where the existence of some definite control of ripple direction is clearly demonstrated, there may still be room for a variety of interpretations as to the position of former

Plate LXIV.

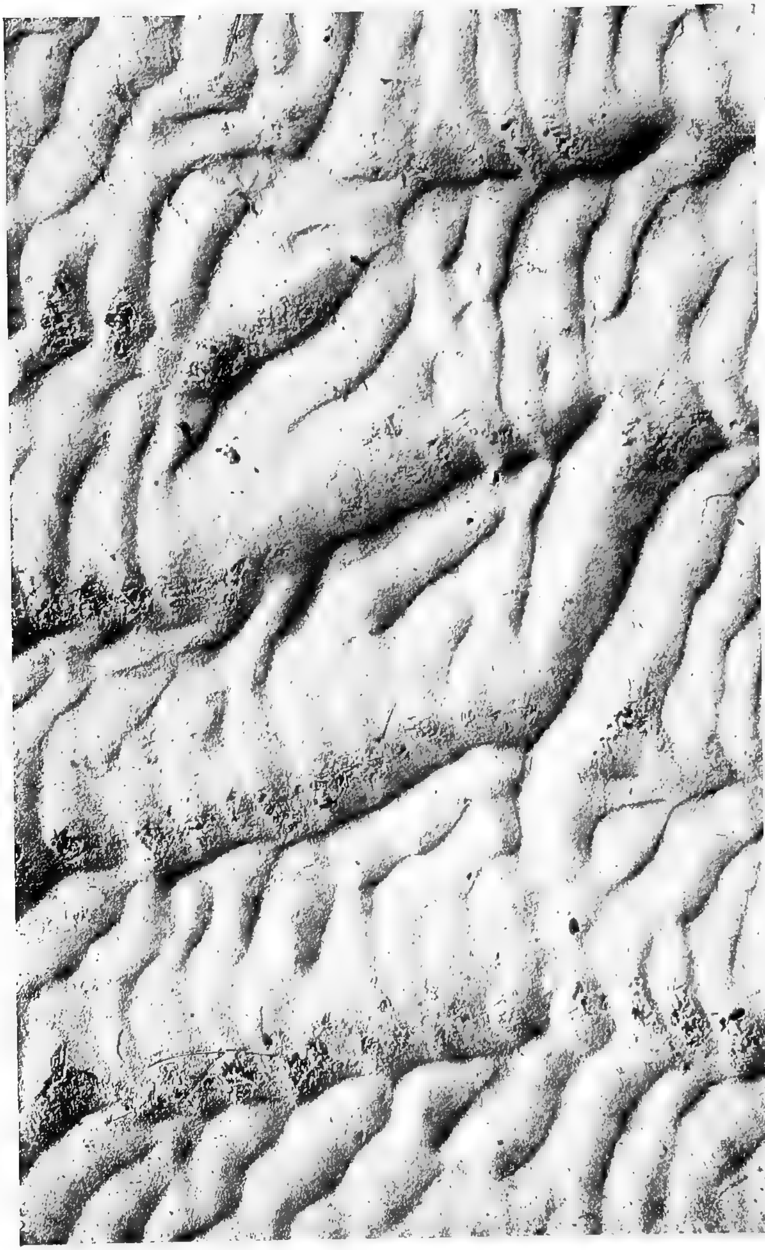


Photo by E. M. Kindle.

Current ripples modified by later oblique wave or current action.

shorelines. An interesting attempt to deduce paleogeographic conditions from a discriminating study of large fossil current ripples will be found in a recent paper by Bucher[86].

The simultaneous action of continuing currents and oscillatory wave motion, as well as the action of intersecting currents or intersecting systems of waves, give rise to a variety of peculiar ripple forms. Thus oscillation ripples may be made slightly asymmetrical by a feeble current, or faint oscillation ripples may be superposed on strongly developed current ripples. Strong oscillatory wave action or secondary current action may give to a series of current ripples the peculiar pattern shown in Plate LXIV, if the waves or current advance obliquely over the earlier formed current ripples. "Interference ripple mark" (Plate LXV) results when two sets of symmetrical ripples are formed by two systems of waves crossing nearly at right angles. The cell-like pattern of some interference ripple marks led Hitchcock to regard them as "tadpole nests." Examples of these and other abnormal ripple types are described and figured by Kindle.

Ripple marks have repeatedly been discussed in connection with the interpretation of fossil ripples found in sedimentary rocks. We need mention but a few of these discussions in the present connection. As early as 1831 Scrope[87] described fossil ripple marks found on slabs of marble, and explained them as due to the oscillatory movements of shallow water. Darwin[88], starting from the very questionable assumption that a great ebb and flow of the tide is essential to the formation of numerous ripples, concluded that the presence of a large number of ripple marks in a geological formation indicated with a considerable degree of probability that the tides of early times rose higher than those of today. Van Hise[89] figures and describes one type of oscillation ripples, and emphasizes their value as criteria for determining the original attitude of steeply inclined strata. Gilbert[90] suggested the possibility of an analogy between ripple marks and vibrations in elastic bodies, basing the suggestion on observations of fossil ripple marks in the Jurassic limestone and Triassic sandstones of Utah.

Spurr[91] shows that where continuous deposition takes place from a current which constantly maintains asymmetrical ripples on the surface over which it flows, the forward movement of the ripples combines with the deposition of heavier and larger

PLATE LXV.

Photo by E. M. Kindle.

Plaster cast of interference ripple mark formed by two systems of waves crossing nearly at right angles.

fragments in the troughs and lighter particles on the crests, to give a peculiar type of false bedding in the resulting formation. Jagger[92] criticized Spurr's conclusions on the ground that his own experiments and observations indicated that ripple marks could not be produced in heterogeneous material; but Spurr[93] met the criticisms with a fuller discussion of the matter in which his original contention is well sustained A short time previously Sorby[94] had described a somewhat similar phenomenon in a paper printed almost exactly half a century after the publication of his first account of ripple marks, already cited. From an examination of the "ripple drift" type of false bedding in rocks Sorby believed one could "ascertain with approximate accuracy, not only the direction of the current and its velocity in feet per second, but also the rate of deposition in fractions of an inch per minute[95]."

The finding of ripple-marked limestone has been the occasion of two lines of reasoning regarding the origin of the rock, both based on the assumption that ripple marks cannot be formed in deep water. According to one argument, the ridges and troughs are not true ripple marks, since limestones are necessarily formed in deep water; the other argument holds such limestones to be necessarily of shallow water origin, because the ridges and troughs are true ripple marks. An example of the former argument may be found in Locke's early report[96] on "waved strata" of Ordovician limestone in southwestern Ohio; while Foerste[97] presents the second point of view in discussing the origin of Ordovician and Silurian beds in this same general region. The frequent occurrence of unusually large ripple marks in limestone has been noted by Gilbert[98], Moore and Hole[99], Cushing[100], Miller[101], Kindle and Taylor[102], Udden[103], Prosser[104], and others, the distance from crest to crest of these ripples varying from one foot to two or three feet in most cases, but reaching a maximum of nearly six feet in an example described by Udden. Wooster[105], Kindle[106], and Udden[107] record the association of ripple marks in limestone with the remains of deep water organisms; while Kindle[108] regards the large size of the ripples as independent evidence of a considerable water depth. Shannon[109] found large ripple marks in limestone associated with sun-cracks, but does not state whether the ripples were of the symmetrical or asymmetrical (current) type. The present writer published in the

Journal of Geology for November, 1916, a review of the literature on ripple marks under the title "Contributions to the Study of Ripplemarks[110]," based on studies made for the present work.

The writer's observations of beaches incline him to the opinion that there is comparatively little chance for the preservation and incorporation in the geological record of ripple marks originally formed on typical beaches. As we have already seen, the beach is a temporary and constantly changing deposit, and while both oscillation and current ripples form on sandy beaches, their subsequent destruction is almost certain, even though streams discharging sediment upon the beach may temporarily bury them. Ripple marks formed on the sea-bottom in the offshore zone stand a better chance for preservation, as also do those on tidal mud flats and sand flats. Under none of these circumstances, however, would the opportunities for preservation seem so good as on river flood plains and deltas. Here ripple marks of both principal types are readily formed in shallow ponds, lakes, and stream channels, and later deposition from spreading flood waters may quietly bury them in places secure from future disturbance. Fossil ripple marks are therefore not to be regarded as an evidence of beach deposits, unless associated with independent evidence of a much more reliable character.

Even where fossil ripple marks have a marine origin, their position furnishes no satisfactory clue to the position of the former shoreline. Both on the beach and in the offshore zone the axes of the ripples may lie at any angle to the shoreline, as has been pointed out in earlier pages. Current ripples with axes at variable angles to the shore are very frequently found in low depressions along the beach. Water from the rising tide or from storm waves entering such depressions at any low point, flows through it developing transverse series of asymmetrical ridges. Oscillation ripples may take any position on the beach, and one occasionally sees there a checkerboard pattern of little hollows and mounds representing two sets of oscillation ripples crossing each other at right angles. In the offshore zone both types of ripples have been observed making high angles with the shoreline.

Regarding the relation existing between size of ripple marks

PLATE LXVI.

Photo by M. C. Dickerson.

Rill marks on a sandy beach.

and depth of water in which they were formed, theoretical considerations based on the nature of current and wave action would seem to compel the following conclusions: (1) Giant current ripples manifestly cannot be produced in extremely shallow water; but aside from this narrow limitation, both large and small current ripples may be formed in either shallow or deep water. (2) Large oscillation ripples cannot be formed in shallow water, for large oscillatory waves are impossible where the depth is small. (3) Both large and small oscillation ripples may be formed in deep water; whether the ripples will be large or small will depend upon a number of factors, among which the length and height of the wave and the depth of the water are most important. The fact that small ripples alone are most commonly found in sandstones, while both large and small ripples occur in limestones, is in accordance with conclusions (2) and (3) above; while the predominance of large ripples in limestones might be expected to follow from the sixth law enunciated by Forel: "Ripples once formed do not experience a change in spacing as a result of diminishing amplitude of oscillation of the water." Large ripples once formed in deep water tend to remain, and so to be preserved by burial, despite later oscillations which would of themselves have produced closely spaced ripples.

RILL MARKS

The water left in the sands of the upper part of the beach after the retreat of the tide or after the dying down of storm waves, often carves tiny drainage channels as it flows back to the sea. These miniature river systems are known as *rill marks*, and are not formed below sealevel. They may, however, be formed on any slope of fine-grained, unconsolidated material from the upper portion of which there is a seepage of water, and hence their presence in consolidated rocks is no proof that the rocks in question represent beach deposits. As is the case with ripple marks, the probability of preservation is not so great when rill marks are formed on beaches as when they are formed elsewhere.

The pattern of rill marks (Plate XLVI) often resembles rather closely that of branching plant stems; indeed, so close is the resemblance that casts of rill marks found in sedimentary rocks

PLATE LXVII.

Photo by E. M. Kindle.

Plaster cast of swash marks left by four successive waves on the sandy shore of Lake Erie.

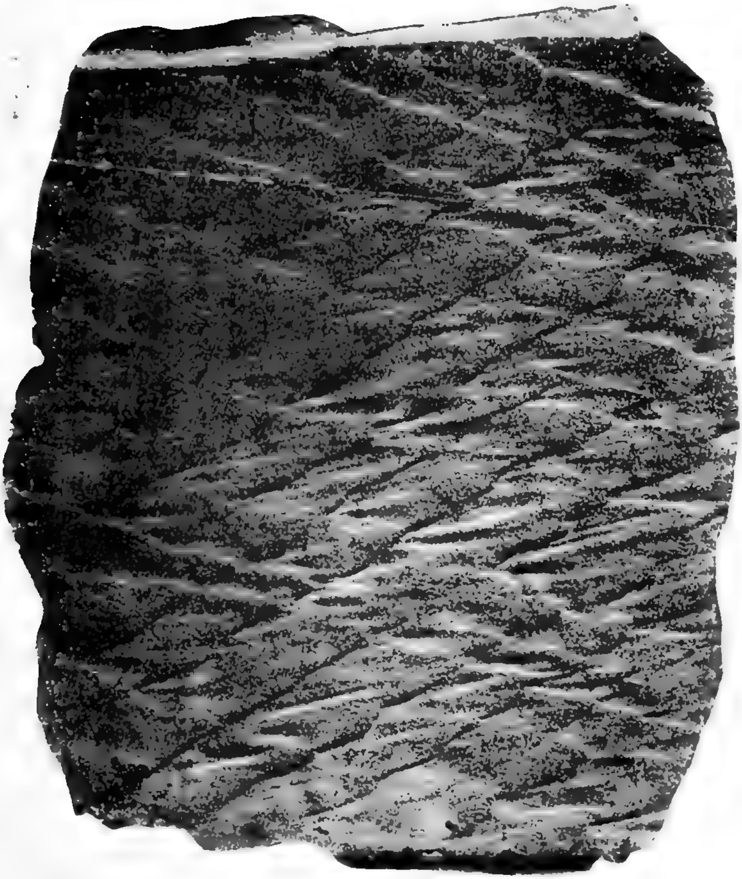
have repeatedly been mistaken for ancient plant remains. In 1873 Nathorst[111] published a paper in which he showed that rill marks and other markings on the strand had been regarded by many authors as phenomena of vegetable origin. I have not seen this paper, but the same idea is briefly presented in the same author's valuable memoir entitled "Om spår af några evertebrerade djur m. m. och deras paleontologiska betydelse[112]," which appeared eight years later. This memoir contains an excellent bibliography of papers treating mechanical markings and the tracks of animals on the shore as vegetable remains, and gave rise to a spirited controversy in which de Saporta[113], Nathorst[114], Gaudry[115], Williamson[116], and others took an active part. Williamson made plaster casts of natural rill marks and showed their identity with many so-called fossil plants. The reader who would follow this phase of the subject further will find additional references to the literature in the works of the authors just cited.

Rill marks of an unusually delicate pattern have been briefly described by Dodge[117] who found them confined to the seaward side of previously formed ripple marks on Winthrop Beach, Massachusetts. Jagger[118] produced artificial rill marks, and described the process of their development. Grabau[119] classes with rill marks those branching distributaries of small streams which debouch upon a beach or other sandy or clayey plain. Rill marks of whatever type present no difficulties as to their origin, while in form they are so simple and unimportant as to require no special discussion.

SWASH MARKS

When a wave breaks at the foot of a gently inclined beach, part of the water glides up the slope in a thin sheet known as the "swash." After the retreat of the swash the greatest advance of its irregular margin is often indicated by a thin, wavy line of fine sand, mica scales, bits of seaweed and other débris, commonly referred to as a "wave mark" (Plate LXVII). Since there are a variety of marks left on sand by wave action, and the present feature is peculiarly a product of the swash, I have given it the name of "swash mark." Although too delicate a feature to attract much attention on the modern shore,

PLATE LXVIII.



Cast of backwash marks (after Williamson).

PLATE LXIX.

Photo by E. M. Kindle.

Plaster cast of backwash marks (after Kindle).

the swash mark is one of the best proofs of beach action usually preserved in sedimentary rocks. When found in the fossil condition swash marks may throw light on other buried shore forms with which they are associated[120].

BACKWASH MARKS

The return flow of the swash down the beach often develops a peculiar criss-cross ridge pattern (Plate LXVIII) in the sand resembling " the overlapping scale-leaves of some Cycadean stem." Williamson[121] regarded similar ridge patterns as the product of intersecting ripple marks trenched by subsequent rills. The illustrations given by him do not suggest such an origin, and I am inclined to regard the phenomena observed by him as identical in origin with the criss-cross pattern which I have observed in process of formation by the backwash. Kindle[122] figures an excellent example of the phenomenon under the title " imbricated wave sculpture" (Plate LXIX), and ascribes it to " the action of very small waves lapping and crossing each other from opposite sides of a miniature spit." It is a matter of common observation that two projecting lobes of the swash are often directed toward each other as they rush up the beach slope, and that the return backwash from the two meet at an angle in their descent. The resultant crossing of currents would be similar to that described by Kindle, and might explain the frequent development of the imbricated pattern on beaches subjected to the action of breaking waves. On the other hand I have observed cases in which it seemed to me the phenomenon was caused by a single backwash current. The thin sheet of water returning down the beach slope appeared to be split into diverging minor currents by every patch of more compact sand or particle of coarser material which impeded its progress, and the crossing of these minor currents resulted in the criss-cross pattern in the sand. Whatever the precise mode of formation, the phenomenon is normally the product of backwash from waves breaking on the beach slope, and may appropriately be called *backwash mark.*

SAND DOMES

When the tide is advancing up the slope of a sandy beach, and the swash from a large wave first sweeps over a portion of the beach previously dry, the disappearance of the water may be accompanied by the appearance of miniature domes or blisters which arise at various points over the area newly subjected to the action of the swash. These domes usually vary from two to eight inches in diameter, and may rise an inch or possibly more above the level surface of the beach. If the curious observer will gently remove one side with a knife blade, he will discover that the dome is hollow as shown in Figure 149, the vertical height of the air chamber corresponding to the height of the dome surface above beach level.

The formation of these *sand domes* may be explained as follows: Before the swash reaches the area in question, the beach

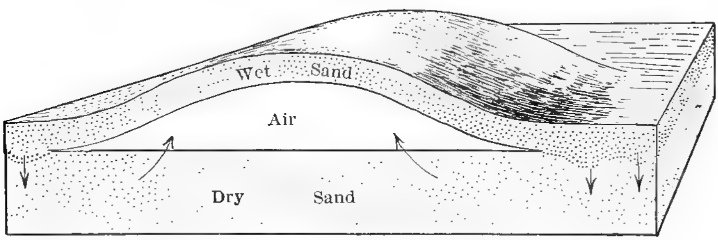
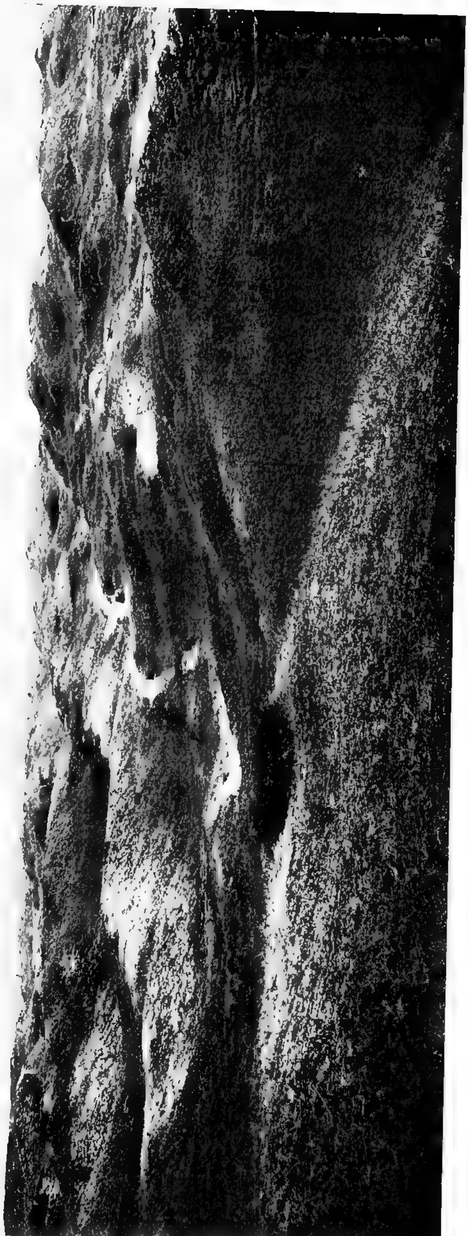


Fig. 149. — Sand dome. Arrows show movement of air as water sinks down from surface.

sands are dry, and air fills the pore spaces between the sand grains. The first advance and retreat of the swash saturates the surface layer of the sand, water replacing air in the pore spaces to a depth of one-fourth or one-half inch. Penetration of the water to greater and greater depths can be accomplished only through expulsion of the air previously occupying the pore spaces. Part of the air escapes directly through the surface film of wet sand, and may be seen bubbling from countless tiny holes before the swash has returned down the beach. In other places the surface film of wet sand is quite air-tight, and is locally raised as a perfect miniature dome by air forced upward through the action of water descending in adjacent areas. Where

PLATE LXX.



Shore dunes of the Holland coast near Katwijk.

by some small mollusc the formation of the dome may be facilitated. It is hardly necessary to remark that the sand domes, which have not to my knowledge been previously described, are very ephemeral features.

SHORE DUNES

The sand dunes, formed from beach sands along the shore, have received much attention in descriptions of shore forms. They are extensively developed along the coast of the Landes in southwestern France, where they attain heights of from 80 to 90 meters in places, cover a belt from 2 to 6 miles in breadth, and have overwhelmed houses and churches causing whole towns to be abandoned by the inhabitants[123]; along the coast of the Netherlands (Plates LXX and LXXI) and Denmark, where they are not so high as in France, but nevertheless serve as an important barrier between the sea and the lowlands reclaimed from tidal waters, attaining a height of 30 meters on the Danish coast; and on the south and east coasts of the Baltic, where they cover broad belts on the Darss foreland and near Swinemünde, and rise to an altitude of 60 meters on the narrow bay bars of the Frische Haff and Kurische Haff[124]. On the Atlantic coast of the United States shore dunes have an extensive development near Provincetown, Massachusetts, and on Cape Canaveral, Florida; while smaller areas on Sandy Hook and other parts of the New Jersey coast[125], near Cape Henry, Virginia[126], and on the offshore bars of the Carolina coast[127] are noted for their dunes. Inasmuch as these dunes are the product of wind action, and are only indirectly related to shore processes, it is not desirable to consider them at length in the present connection. The only dunes which have special interest for the student of shore processes are those occurring in the form of parallel ridges on a beach plain. These " dune ridges " have already been fully discussed in Chapter IX.

The student desiring to pursue further the study of dunes should consult the early work of Brémontier[128] bearing the title "Mémoire sur les Dunes."

Solger's "Dünenbuch[129]," includes a treatment of shore dunes, and Sokolów, in his important work entitled "Die Dünen: Bildung, Entwickelung, und innerer Bau[130]," discusses sand

PLATE LXXI.

Shore dunes near Scheveningen, Holland.

PLATE LXXII.

Dune of barchane form overwhelming trees on the Provincelands of Cape Cod. The top branches of a tree protrude from the crest of the dune.

dunes of all types and gives copious references to the literature of the subject. He reaches the conclusion that over 90 per cent of the shore dunes of Europe occur on coasts which are subsiding, or which at least are being undermined by wave attack; and interprets this to mean that the undermining action constantly uncovers fresh supplies of sand and hinders the growth of vegetation which might protect the sand from wind action, whereas on a rising coast sand deposits may be raised above the reach of the waves and be replaced by clay or other sand-free sediments[131].

If Sokolów's conclusion and interpretation were valid, the presence or absence of shore dunes would become a matter of much importance in determining past changes of level. Unfortunately, the criteria accepted by this author as satisfactory proofs of land sinking would probably result in the classification of 90 per cent of all the coasts of the world as sinking coasts; whereupon the occurrence of 90 per cent of the dunes upon such coasts would lose significance. Neither can we agree that retrograding coasts necessarily favor, and prograding coasts hinder, dune formation. The almost complete absence of shore dunes on some of the European and American coasts suffering most from wave attack, and the magnificant development of dunes on such prograding shores as those of the Darss, Swinemünde, and Cape Canaveral, point to a different interpretation. The development of shore dunes depends upon a number of variable factors, among which are the direction of the wind (offshore or onshore), the rapidity with which débris is supplied to the shore, the size of the débris particles, the nature of the climate, and the stage of development attained by the shoreline. It may be doubted whether very slow changes of level constitute a factor of importance. In any case, it would seem that a retrograding shoreline, along which more material is being taken from the land than is added to it, would present conditions unfavorable to the extensive accumulations of shore dunes; whereas, it is certain that the dunes of the Darss, Swinemünde, and Canaveral, and probable that those of the Landes, owe both their formation and their preservation to the prograding of sandy shores.

PLATE LXXIII.

Shore dunes near Cape Henry, Virginia, migrating inland over the forest.

RÉSUMÉ

In the present chapter we have turned our attention to those minor forms of the shore zone which have no great significance in the general history of the shore cycle, but which nevertheless appeal to the interest of every observer who studies the meeting place of land and water with an inquisitive mind. It has been shown that the triangular cusps of sand or gravel built by waves upon the beach have given rise to much discussion and to several theories of origin. These theories we have examined and criticized in the light of new evidence as to the distribution and characters of the cusps. The low and ball of sandy shores have been briefly treated, and the puzzling problem of their origin indicated by citations from different observers. We have examined some of the rather abundant literature relating to the interesting phenomena of ripple marks, and have noted the value which these forms have to the geologist who must interpret the origin and structure of sedimentary rocks. Rill marks, swash marks, and the marks produced by the backwash in turn received brief attention; while the curious but very temporary sand domes have been described and their origin explained. Finally the interesting sand dunes occurring on the shore have been mentioned, and suggestions offered as to where the student may find elaborate discussions of these forms, which do not properly lie within the province of a book devoted to shore processes and shoreline development.

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